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North Atlantic Deep Water Formation

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Proceedings of a workshop held at Lamont-Doherty Geological Observatory Palisades, New York June 4-5, 1984

NASA

NASA Conference Publication 2367

North Atlantic Deep Water Formation

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Proceedings of a workshop held at
Lamont-Doherty Geological Observatory
Palisades, New York
June 4–5, 1984

NASA
National Aeronautics
and Space Administration

Scientific and Technical
Information Branch

1985

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PREFACE

A miniworkshop on North Atlantic Deep Water Formation was held on June 4-5, 1984 at Lamont-Doherty Geological Observatory, based on financial support from the Goddard Space Flight Center Director's Discretionary Fund. The workshop is expected to be the first in a series on key problem areas in the earth sciences. The objective is to contribute to improved definition of a scientific approach for understanding each problem, including data and measurement requirements. This should also help define contributions that NASA could make as part of its Global Habitability program.

The workshop approach is to bring together scientists from different disciplines that might not usually interact, particularly researchers who think across disciplinary boundaries. Prior to the workshop a focused question is identified within the broader scientific problem area.

The broad issue of concern for the first workshop was: How will ocean circulation respond to climate change at the ocean surface, and what are the effects of ocean variability on climate? The focused question which the workshop addressed was: What controls the rate of deep water formation in the North Atlantic Ocean and what repercussions would there be from changes in this rate?

North Atlantic Deep Water (NADW) forms near the end of at least some winters, after cooling and evaporation at the ocean surface have created water of sufficient density that it sinks to the ocean bottom somewhere in the Greenland/Norwegian/Labrador Sea region. Eventually this deep water spreads through and ventilates the deep global ocean; the process of NADW formation is the principal mechanism causing the exchange of heat, nutrients, CO_2, and other trace substances between the deep ocean and the surface layers and atmosphere. It is thus apparent that any major changes in deep water formation could have important implications for climate, ocean productivity, atmospheric CO_2 and related issues.

It can be said that deep water forms in the North Atlantic because it is the saltiest of the high latitude ocean basins. But why is it the saltiest ocean and how subject is this to change? Deep water formation itself probably helps create the high salinity by drawing north warm waters which are subject to high rates of evaporation. Is such a self sustaining mechanism for creating dense deep water stable to climate fluctuations, natural or man-made, at the ocean surface? Are other source regions for North Atlantic water, such as the Arctic and Mediterranean Seas, subject to climate fluctuations sufficient to influence the rate of deep water formation in the North Atlantic? Since conditioning of the surface water which sinks to form NADW occurs in the upper layer of the ocean, it seems possible that climate changes at the ocean surface could modify deep water formation on short time scales.

Evidence for rapid changes of climate in the North Atlantic region, in fact, has been found in studies of isotopic abundances in Greenland ice cores by Dansgaard, Oeschger and others, in studies of North Atlantic sediment cores by Ruddiman, McIntyre and others, and in studies of pollen records throughout Europe. Several climate fluctuations from near glacial to near interglacial conditions appear to have occurred in periods not longer than a few decades. Recent measurements by Oeschger and colleagues of the CO_2 content of a Greenland ice core suggest similarly rapid fluctuations of atmospheric CO_2. If more complete analysis shows these apparent changes in atmospheric CO_2 to be real, they may be difficult to explain other than through changes in ocean circulation, because of the limited magnitude of CO_2 reservoirs on land.

Oeschger interprets the rapid climate changes as an indication that the climate system may shift between two quasi-stable states. Broecker et al. suggest that these climate states may be one in which NADW formation is 'on' and one in which it is 'off'. Climate models predict large increases in temperature and precipitation at high latitudes in the coming century as a result of the greenhouse effect of increasing atmospheric CO_2 and trace gases. Could this climate change shut off or reduce deep water formation? If so, the North Atlantic and European regions may actually become colder while the global climate warms. In any case, the paleoclimate data indicates that current perceptions of long-range climate issues are probably too simple. In particular: 1) climate changes can happen more suddenly than presently anticipated, and 2) the geographical distribution of climate changes may be much more complex than suggested by current climate models, which assume that the ocean will continue to operate in the future in the same way as at present.

Wallace Broecker served as scientific chairman for the workshop, James Hansen as co-organizer and Theodore Bennett as recorder. Participants are listed in an Appendix.

SUMMARY

Background. Interactions between climate and the ocean are believed to be important on all climate time scales. Recently it has become increasingly clear that the ocean may play a major role in determining climate trends on decade to century time scales.

The effect of ocean heat capacity as a thermal buffer on climate has been emphasized by Charney (1979) and others. Recent empirical paleoclimate evidence, especially the ice core data of Dansgaard, Oeschger and their colleagues, indicates that climate and perhaps also atmospheric carbon dioxide have undergone large fluctuations on time scales of 10-100 years (see, for example, Dansgaard et al., 1984; Oeschger et al., 1984). Although the explanation for the paleoclimate fluctuations may be complex, their existence indicates that we can not assume that ocean circulation and mixing will operate in a fixed way over decadal time scales.

On the basis of considerations such as the large magnitude of climate fluctuations in the North Atlantic region, the unique role of North Atlantic deep water in the world ocean circulation, and evidence from oceanographic measurements and ocean tracer data that the deep water formation process is highly episodic, we chose to focus this workshop on the question: What controls the rate of deep water formation in the North Atlantic Ocean and what repercussions would there be from changes in this rate?

Presentations. Presentations were organized in three half-day sessions on oceanography, paleoclimate and modeling. Extended abstracts and key figures from each of these presentations are included in the next section, as the bulk of this report.

Recommendations. Strategies for improving our understanding of deep water formation were discussed in the final half-day session. Recommendations are presented in the final section below, the principal aspects being the need for the following studies:

Oceanographic measurements:

1) Hydrographic cruises each spring into the Norwegian, Greenland and Labrador Seas to determine the amount and properties of NADW formed each winter for a 10 year period.

2) Monitoring of the meteorological conditions before and during deep water formation to allow investigation of air-sea interaction events, such as the role of cyclogenesis in deep water formation, and monitoring of the sea ice distribution.

3) Monitoring over the 10 year period of the currents and T/S properties of the outflows across the Greenland-Iceland-Faeroes-Scotland ridges into the North Atlantic Ocean.

4) It is also desirable to monitor over this 10 year period currents and T/S properties of the inflows of source waters to the deep water formation regions from the Arctic, the subtropical gyre, and intermediate waters from the Labrador, Irminger and Mediterranean Seas. An alternative is to define a sampling strategy for transient tracers and nutrients with a frequency adequate to define the contributions from different source waters.

5) Development of the required improved capabilities for moored and drifting buoys, which can communicate results via satellite.

6) Release of a network of XBTs into regions of developing deep convection.

Paleoclimate data:

1) Exploration for marshes and lakes with pollen records extending back at least 40,000 years, and detailed studies of such records.

2) Retrieval and study of ocean cores from areas of high deposition rate (i.e., >6 cm/10^3 years) in the North Atlantic.

3) Retrieval and study of ice cores from the Crete site in Greenland (site chosen by NSF for the next drilling effort).

4) Extension of the tree ring record to all geographical regions around the North Atlantic basin.

Satellite data:

1) Monitoring of quantities required to infer air-sea fluxes, specifically radiation budget, sea state (for surface winds) and sea ice.

2) Repeated altimetry of the North Atlantic to help monitor the large scale circulation and to detect and monitor the surface water depression associated with deep water formation events.

3) Development and use of greater capabilities for high precision measurements from moored and free-floating buoys with satellite data relay.

Modeling:

1) High resolution modeling of the North Atlantic and Arctic Ocean.

2) Fully coupled ocean/sea ice/atmosphere modeling, which allows analysis of the factors influencing deep water formation.

3) Testing of models against paleoclimate data, as well as against today's climate.

REVIEW OF NORTH ATLANTIC SOURCE WATERS

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North Atlantic Deep Water (NADW) ventilates the deep World Ocean. It not only carries relatively well-oxygenated waters, but also other substances derived from recent sea-surface exchanges. There are five regional sources for NADW (Fig. 1): 1) derivatives of the salty Mediterranean Sea outflow, 2) products of open-ocean convection in the Labrador Sea, 3) Iceland-Scotland Overflow Water from the Norwegian Sea - salty by virtue of mixing with saline water near the sills, 4) Denmark Strait Overflow Water from the Iceland and Greenland Seas - which retains a high-density, relatively low-salinity signal, and 5) remnants of deep water from the Antarctic circumpolar region - freshest of the bottom waters. Despite the differences of characteristics of the source waters, the NADW is relatively uniform (e.g., cf. Fig. 1).

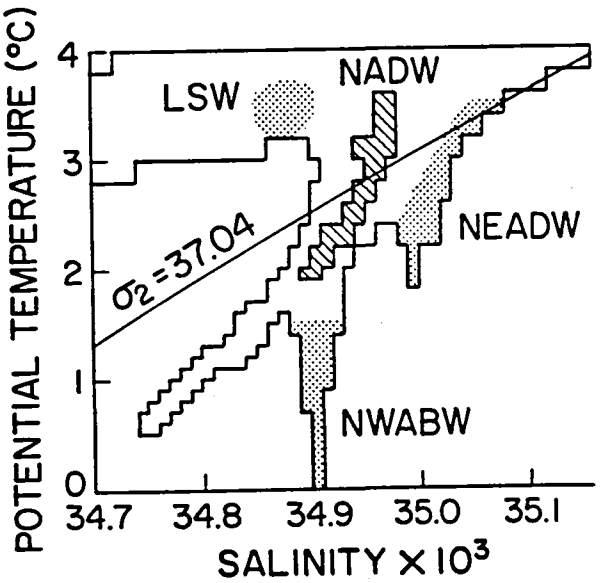

Fig. 1. Potential temperature - salinity diagram for North Atlantic deep waters colder than 4°C, adapted from Worthington and Wright (1970). The solid line encloses all of Worthington and Wright's θ-S classes containing at least 100 km^3 of cold deep water. The hatched θ-S classes are the largest classes enclosing 51% of the total deep water volume, i.e., the principal volumetric mode (NADW: North Atlantic Deep Water). Regional water mass nomenclature is introduced by shading near the appropriate θ-S ranges: LSW (Labrador Sea Water), NEADW (Northeast Atlantic Deep Water, fed by the Iceland-Scotland overflow), and NWABW (Northwest Atlantic Bottom Water, fed by the Denmark Strait overflow). The waters of the cold, fresh "arm" are derived from the Antarctic. The Mediterranean Sea outflow waters are too warm and salty to appear distinctly within these θ-S ranges. The θ-S correlation at a single northern North Atlantic location may show influences of each of these deep water masses, but would not pass through all the shaded areas. (From Swift, 1984b)

Because the formation of each of the five source waters may be viewed as a response to a complex series of events, it is difficult to examine the sensitivity of NADW to environmental fluctuations. It is known that the deep northern North Atlantic is relatively closely coupled to the sea surface in the Greenland and Iceland seas. The most recent studies indicate a minimum response time of only two years between the introduction of a passive signal north of Iceland and its appearance in the deep northwestern Atlantic (Livingston, Swift, and Ostlund, 1984). Major deep ocean responses are possible: between 1972 and 1981 the northern North Atlantic (north of 50°N) freshened by about 0.02°/oo and cooled by about 0.15°C below 2000 meters (Fig. 2; see also Brewer et al., 1983, and Swift, 1984a). The deep water salinities responded to a freshening during the 1970's of the upper layers in the deep water mass formation regions. This upper-layer freshening may in turn be related to shifts in the large-scale atmospheric forcing over the northern North Atlantic and Greenland, Iceland, and Norwegian seas. At the present time the overflow transports are only approximately determined, and long-term variations in deep water formation rates, for example, have not yet been measured.

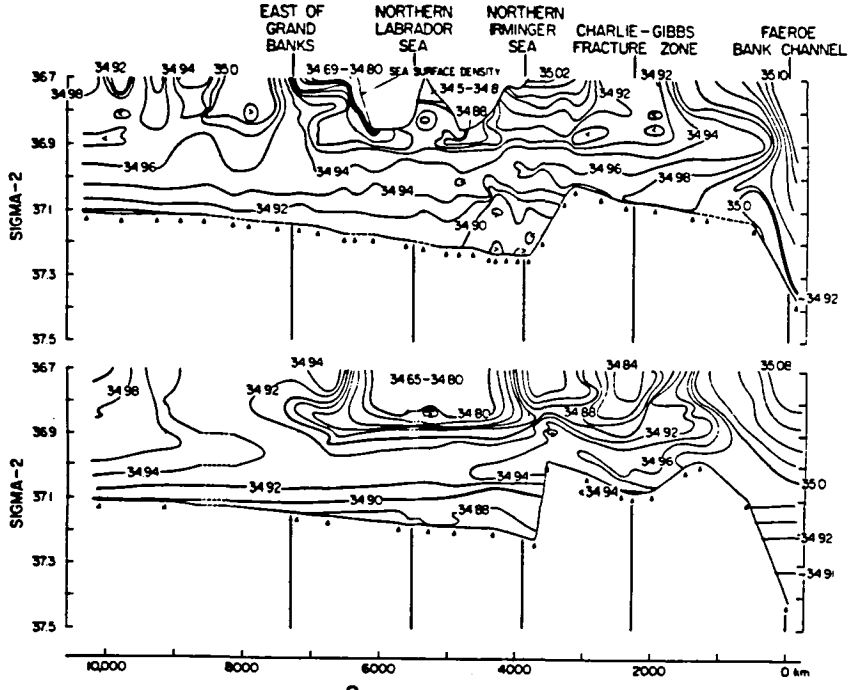

Fig. 2. Sections of salinity (X 10^3) versus σ_2 on a long, winding path from the Faeroe Bank Channel to the waters off the U.S. east coast (location shown in Figure 3). The upper section is drawn from a composite of historical data (Chain, 1960 and 1972; Crawford, 1960 and 1965; Erika Dan, 1962; Atlantis, 1964; Hudson, 1967; Knorr, 1972 and 1976) and the lower section is drawn from April-October 1981 TTO/NAS data. The only post-1972 data in the upper section is a single Knorr 1976 station in the Faeroe-Shetland Channel. The 1981 data from layers where $\sigma_2 > 36.9$ show an average freshening of ca. 0.02 x 10^{-3}, and since layer densities are about the same, there was also a cooling of ca. 0.15°C. The principal deep layers are LSW, near $\sigma_2 = 36.9$, NEADW or Iceland-Scotland overflow water, near $\sigma_2 = 37.0$, and NWABW or Denmark Strait overflow water, below 37.05 in the northwest Atlantic. All displayed the θ-S shift. See Swift (1984a) for a more thorough discussion of this figure and related results.

Another issue of interest is why the North Atlantic produces new deep water while the North Pacific does not. Many factors have been proposed, such as effects of basin geometry and evaporation/precipitation/run-off differences. The North Atlantic also includes two principal Mediterranean seas: the Mediterranean Sea and the combined Arctic Ocean, Greenland/Iceland/Norwegian Sea system. The first is an evaporation basin, and it supplies dense salty water to the North Atlantic. The high-salinity waters extend north into the second Mediterranean region, where they are cooled to near-freezing temperatures and overturn to provide the dense overflow waters (for example, see Aagaard, Swift and Carmack, 1984). To the extent that this is true, it is the presence of these Mediterranean seas which triggers the present-day thermohaline circulation and makes NADW the primary ventilator of the deep World Ocean.

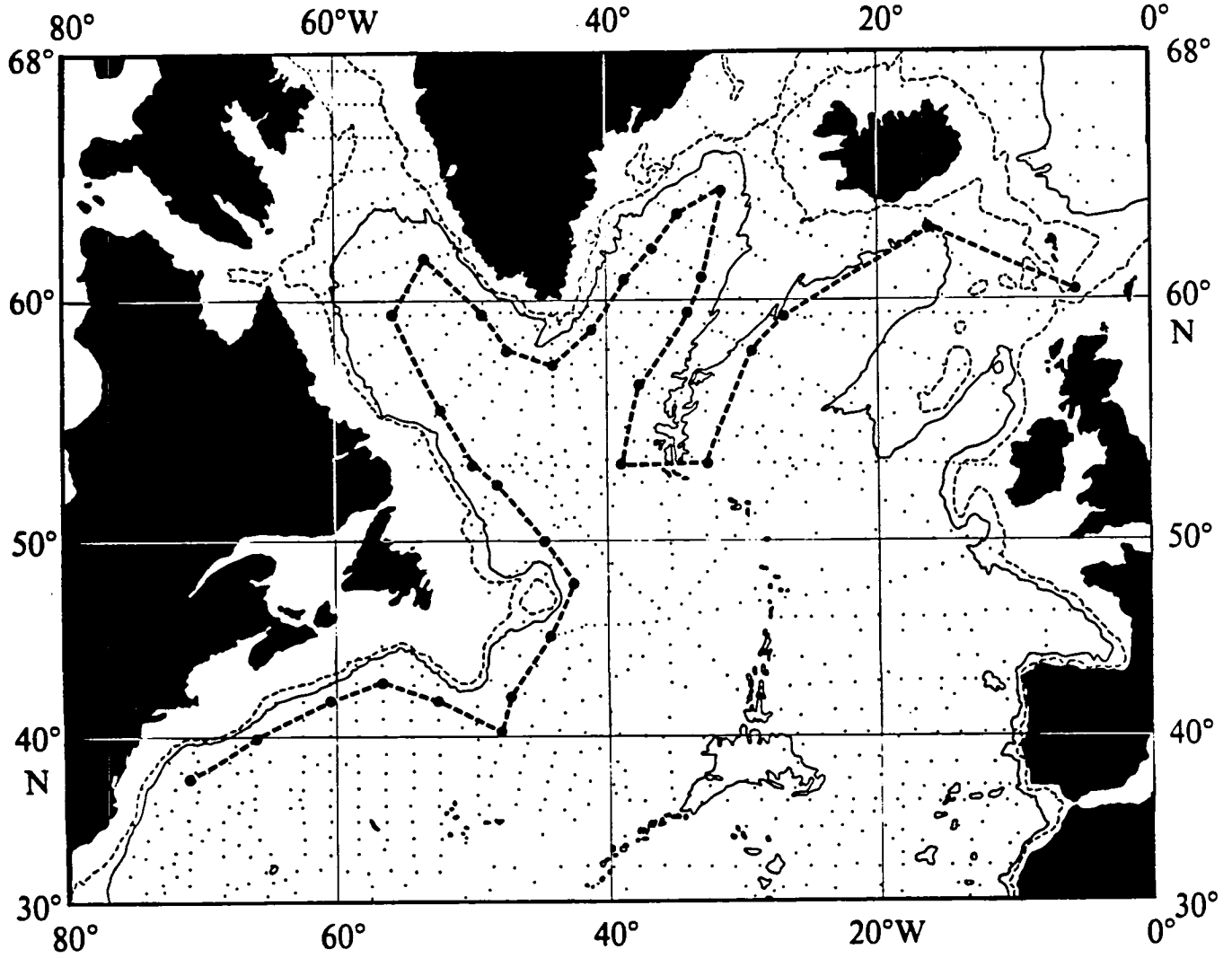

Fig. 3. Location of the sections in Figure 2. The 500 m and 2000 m isobaths are adapted from General Bathymetric Chart of the Oceans (GEBCO) numbers 5.04 (1978 edition) and 5.08 (1982 edition).

DEEP WATER FORMATION

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In this talk I attempted to cover, from a theoretician's point of view, many of the features of deep water formation, specifically as they apply to North Atlantic deep water.

The talk began by noting that the topic of the meeting was considered highly germane by the WOCE (World Ocean Circulation Experiment); their specific objective 4 related specifically to water mass formation. [For a discussion of WOCE, see Nowlin (1984).] First, some simple arguments on plumes of dense water and filling boxes were given. What determines the time for a large-scale environment to be modified by the injection of dense water at its edge is the mass flux, not the buoyancy flux. However, it is the denser buoyancy flux, when there are several competing plumes (e.g. the Mediterranean outflow versus the Denmark Strait outflow) that determines which plume will provide the bottom water for that ocean basin.

It was noted that the 'obvious' laboratory experiment (rotate a pie-shaped annulus, and heat/cool it on the surface) had never been performed. Thus, to some extent our belief that deep convection is somehow automatic at high latitudes to close off some ill-defined meridional circulation has never been tested.

A summary of what we believe to be true about deep convection was given, taken largely from Killworth (1983). The two fundamental formation mechanisms are shown in Fig. 1. Of the two, it is open-ocean convection which forms the water which supplies the Denmark Strait overflow -- in all likelihood, as formation in the Greenland Sea remains stubbornly unobserved. But it is the slope convection which finally creates North Atlantic deep water, following the Denmark Strait overspill.

The question was then asked: Why does deep water form in the Atlantic anyway? As an illustration, the flow of a dense tongue or plume of water over a parabolic depth profile was discussed. It was shown that the asymptotic depth of descent of such a plume (assumed to emerge from a very deep reservoir) is roughly proportional to the product Cg/f^2 (C=curvature) times the fractional buoyancy difference between plume and surrounding waters. This fraction is minute for the Denmark Strait (0.007) but quite large for Gibraltar (0.11). This suggests that Denmark Strait overflow should not form bottom water, but should simply turn and flow horizontally along depth contours at a shallow depth, with forces balanced between Coriolis upward and gravity downward.

However, recent laboratory and theoretical work by Griffiths and Linden (1982) and Killworth, Paldor and Stern (1984) suggests that such plumes are highly unstable, and form large-scale lenses of dense water. Positive curvature of the bottom is a slightly stabilizing feature here. Since rotation is not dominant in the Denmark Strait, there is little small-scale mixing, so that any other frictional mechanism will provide an extra component to the triangle of forces acting on the plume and allow the plume/lenses to slowly descend the slope to the bottom. Conversely, the Gibraltar outflow is dominated by rotation, and the resulting small-scale mixing diffuses the plume away before it can reach the bottom, as Smith's (1975) model shows.

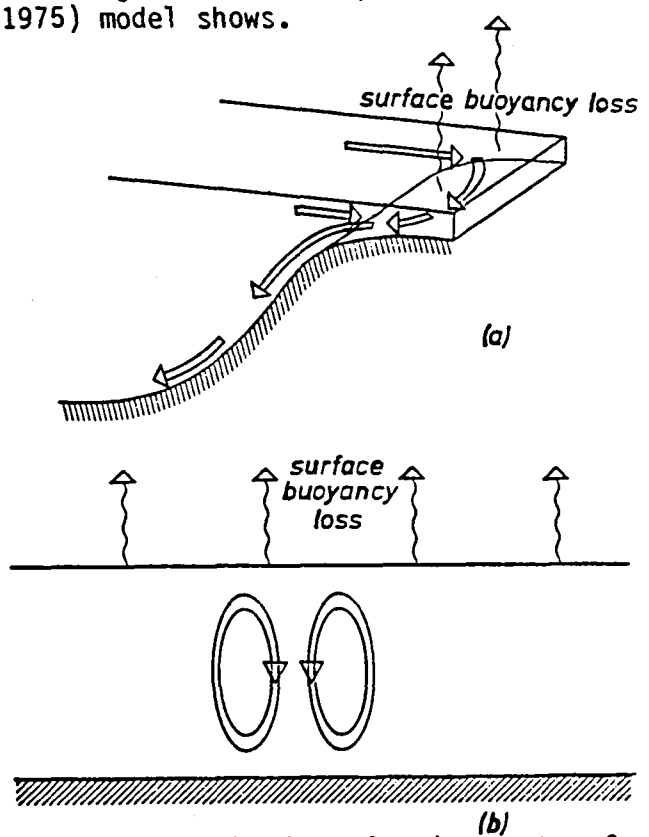

Fig. 1. Two basic mechanisms for deep water formation.

Once it has reached the bottom, dense water is believed to spread largely in western boundary layers. Models of this phenomenon reproduce the feature fairly well, but the vertical depth of penetration of surface injected tracers like tritium is badly reproduced. The same is true for active tracers like potential vorticity.

The reason for this failure (omitted due to time constraints in the talk) seems to be the poor way that models handle the surface mixed layer (and hence, by extension, deep convection). The usual belief is that tracers both passive and active are injected into deeper water by vertical Ekman pumping at the base of the mixed layer. A simple order-of-magnitude estimate, however, suggests that horizontal advection through the sloping mixed layer/stable fluid interface injects 2 to 5 times as much tracer as does vertical motion.

Unfortunately, large-scale models do not include an accurate representation of the deep surface mixed layer (indeed, most cannot resolve the sorts of 20 km chimney we suspect to be responsible for deep convection in the Greenland Sea). Nor do they usually include seasonal effects, which Woods (1984) has demonstrated can dominate the production of both passive and active tracers in the subsurface ocean.

Particle trajectories are only now beginning to be included as diagnostics in large-scale ocean models, and these will be vital if we are to understand where the deep water in the Atlantic finally ends up. Even fairly simple models (like a 'thermocline equation' model I showed) can have remarkably convoluted particle paths. Fig. 2 shows a typical example. On the left are the horizontal velocity vectors (note change of scale) for a two-level ocean, which allows vertical velocities between the levels, which are of constant depth. The larger diagram shows the trajectory of a particle injected into the 4000 km square basin at the base of the subtropical Ekman layer at the point marked with a circle. The particle is then advected three-dimensionally with the steady flow on the left. Firm lines denote downward flow, dashed lines upward flow. The route which allows ventilation of the subtropical gyre involves recurring visits to the northern subpolar gyre and sinking on the northern boundary. Notice how little of the deep subtropics is actually ventilated. (When the crosses, placed every year, are far apart, the particle is in the upper level; when close, in the lower level).

As with every other branch of the discipline, a great deal more work is still needed!

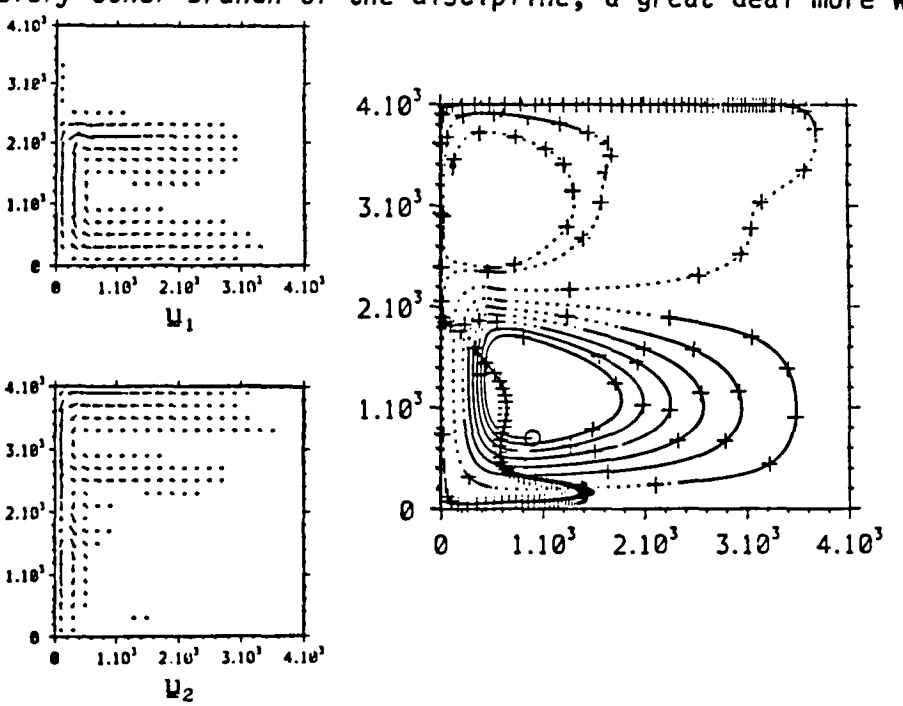

Fig. 2. Horizontal velocity vectors in a two-level ocean model (left figures) and trajectory of a particle injected at the base of the Ekman layer at the point marked by a circle (right figure). Small horizontal arrows show 10 cm s^{-1} in upper level, 1 cm s^{-1} in lower level. Coordinates show distance in km.

TEMPORAL AND SPATIAL SCALES OF LABRADOR SEA WATER FORMATION

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Labrador Sea Water is an intermediate water found at the same density and depth range in the North Atlantic as the Mediterranean water. It is formed by convection from the sea surface to depths greater than 2 km in winter in the Western Labrador Sea.

The processes leading to deep convection begin with the formation of a 200 km scale cyclonic circulation about denser than average upper layer water in the Western Labrador Sea (Figure 1). This circulation pattern is hypothesized to be driven by an ocean/atmosphere heat exchange that has its maximum in this region (Clarke and Gascard, 1983).

By early March, if deep convection is taking place, one sees that this body of denser upper waters penetrates to the top of the deep temperature/salinity maximum marking the core of the North Atlantic Deep Water (Figure 2). We note that the horizontal scale of this body is still 100-200 km normal to the coastline.

If we examine the details of the kinematics taking place within the cyclonic circulation, we find that there are two scales of motion, a mesoscale structure with a radius of 30-40 km and an eddy scale feature that is the order of 10-20 km (Figure 3). The mesoscale structures have a rotational period of the order of 10 days with the eddy scale periods 1 to 2 days. It is hypothesized that an instability of the mesoscale features forms the eddy scale features. The densest and most homogeneous water columns were found within the eddy scale features. The features attain their anti-cyclonic circulation by greatly depressing the pycnocline found at the top of the North Atlantic Deep Water. By the end of March, 1976, the homogenous water columns in the center of the eddies reach depths greater than 2.2 km. The pycnocline outside of the eddies and mesoscale was elevated to 1.2 km.

The actual vigorous deep convection which takes place within these eddies, occurs on short time and space scales in response to strong atmospheric forcing. Our vertical current meter results (Gascard and Clarke, 1983) suggest space scales of a kilometer or less and the same time scale as that of the storm causing the vigorous convection (12 hours). The dynamics of the eddy also responded to these periods of intense downwelling.

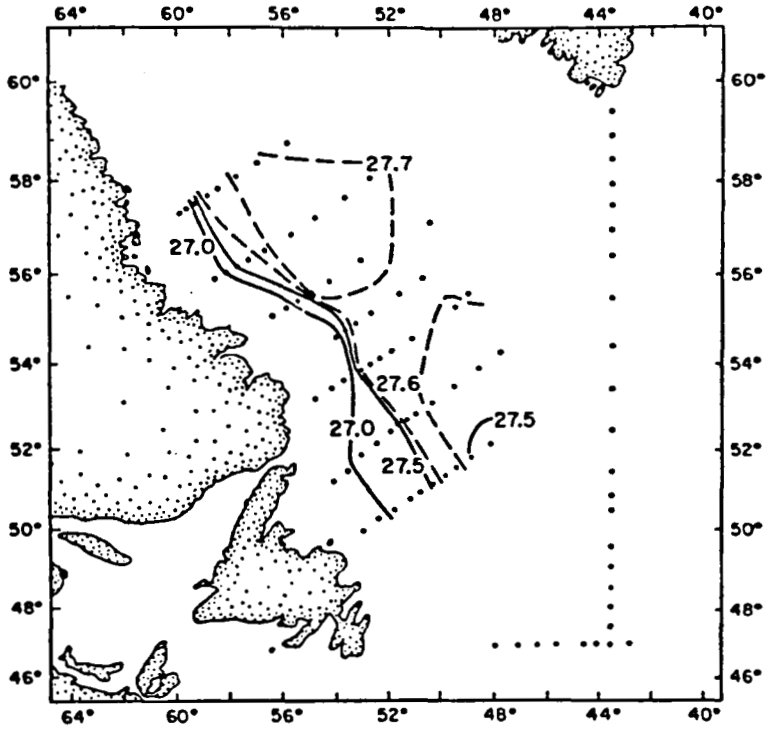

Fig. 1. Potential Density at 100 dbars in the Western Labrador Sea during February, 1978.

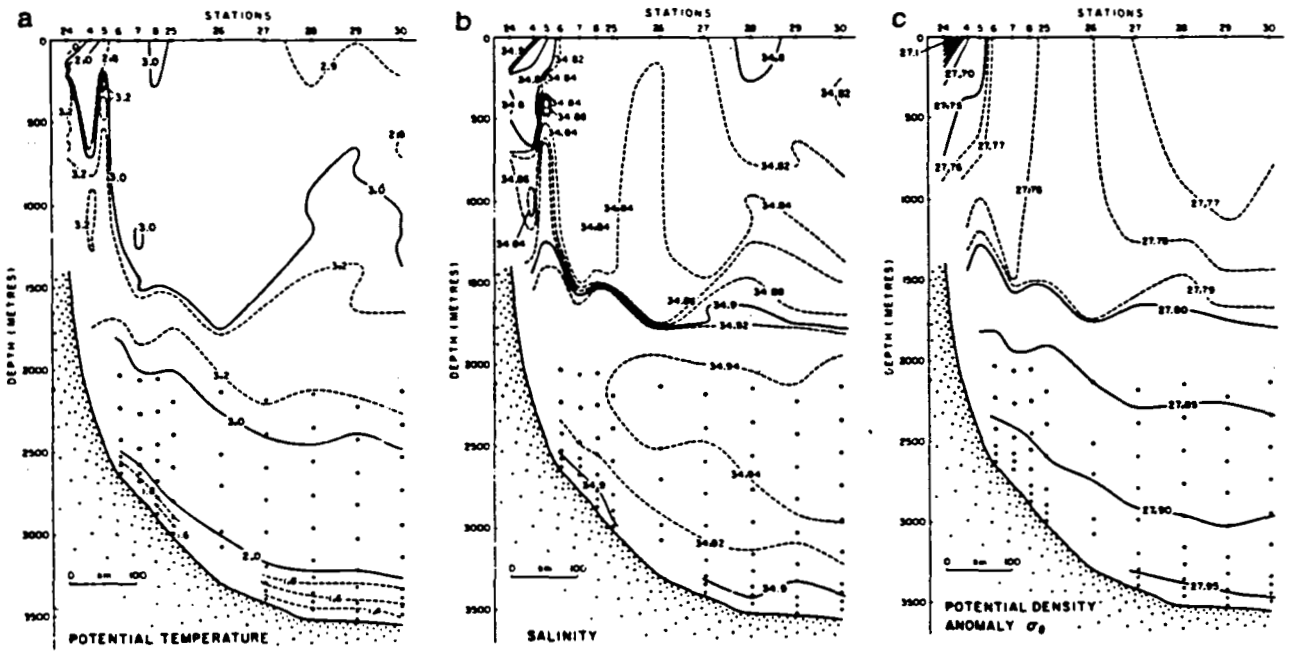

Fig. 2. (a) Potential temperature, (b) Salinity, and (c) Potential density along a section normal to the Labrador slope.

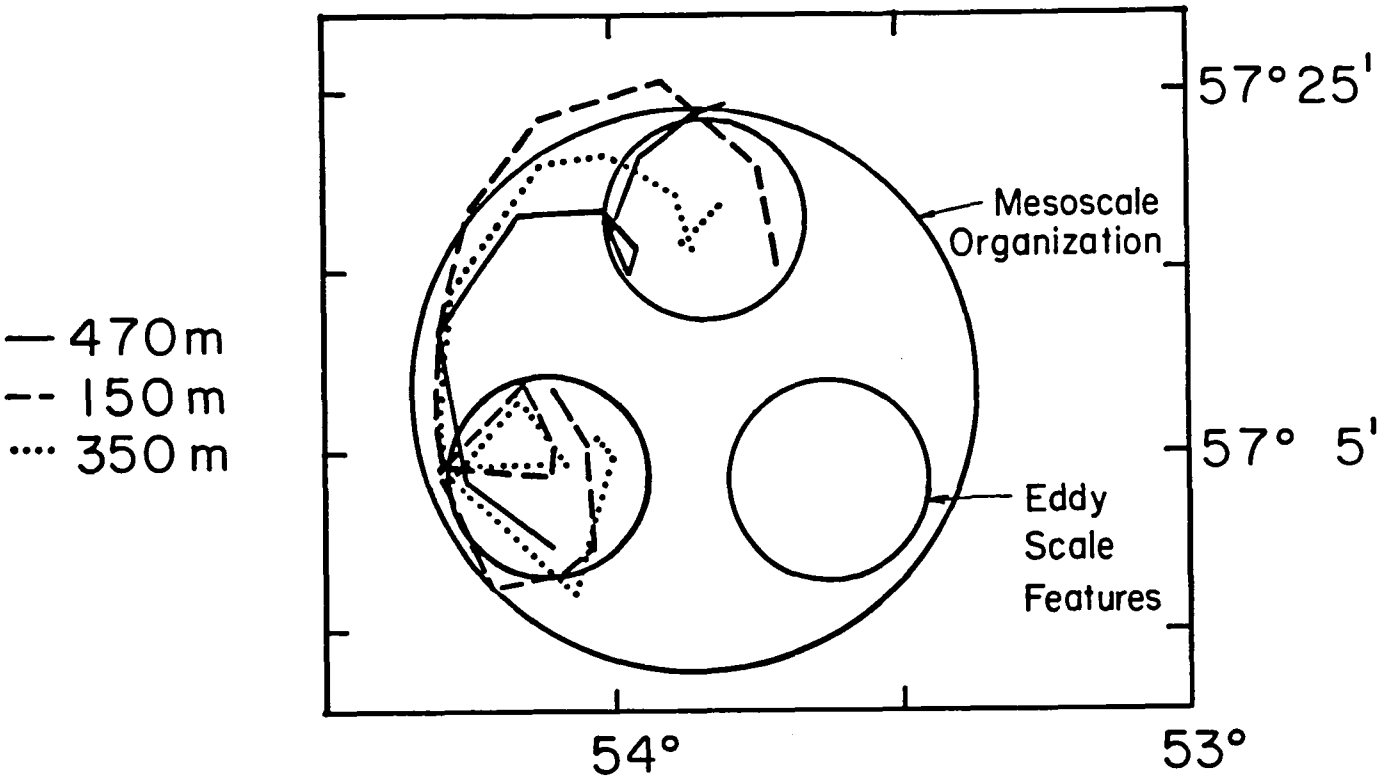

Fig. 3. Conceptual model of how the eddy scale features are related to mesoscale features. The tracks of neutrally buoyant vertical current meters at depths of 150 m to 470 m are superimposed.

These three scales of organization (cyclonic gyre, mesoscale and eddy scale) seen in the Labrador Sea were also observed during Mediterranean Water formation in the Western Mediterranean. There, convection extends to the bottom and both the mesoscale and eddy scale circulations were cyclonic rather than anti-cyclonic. In the Mediterranean Sea, the structures break up quickly (over a few days) once the strong forcing is turned off and once convection has reached the bottom. In the Labrador Sea, the eddy scale features showed no sign of breaking up over the period we were studying them, even though the strong forcing was turned off. Work on "meddies", lenses of anomalous water in the ocean at depth, would suggest that the Labrador Sea Water eddies would be long-lived (several years) provided they were not destroyed through contact with the bottom or by strong shears.

It is also interesting that in the winter of 1978, which was dominated by SE winds rather than NW winds, there was no deep convection. The cyclonic gyre did form; however, there was weak evidence of mesoscale and eddy scale features. This might suggest that the convection is a component of the generation mechanisms for these features.

What can be done from space to monitor these processes? Surface temperature differences within such an area are on the order of 0.1 °C and thus are unlikely to be measurable. Satellite tracked T and S chains are a possibility. Surface buoys do remain in the region for periods of months; the difficulty is the accuracy of the temperature and salinity measurements which would have to be maintained at a variety of levels down to 2.5 km. These accuracies need to be 0.01 °C and 10 ppm in salinity. The mesoscale features may be detectable from surface velocity measurements.

Water types had usually been thought of as invariant by oceanographers. Oceanographic surveys in the 1960's in the Western North Atlantic identified Labrador Sea Water as being 3.4°C, 34.9°/oo (Lazier, 1973). In 1976, we observed Labrador Sea Water being formed at the same density but at 2.9°C and 34.84°/oo. One sees that there has been a considerable range of T and S; however, LSW always appears as a salinity minimum at σ_θ = 27.78. This density restraint is applied by the presence of the higher salinity North Atlantic Deep Water beneath the LSW. For example, to increase the density of a 2 km column of LSW by 0.01 Kg/m^3 would require the heat and fresh water loss equivalent to evaporating 40 cm of water from the sea surface. This would require strong cooling conditions to continue for another month. Since formation of LSW itself only occurs during intense winters, such an occurrence would be an exceptionally intense and prolonged winter cooling. 0.01 Kg/m^3 is about the detection limit for changes in the density of LSW from decade to decade.

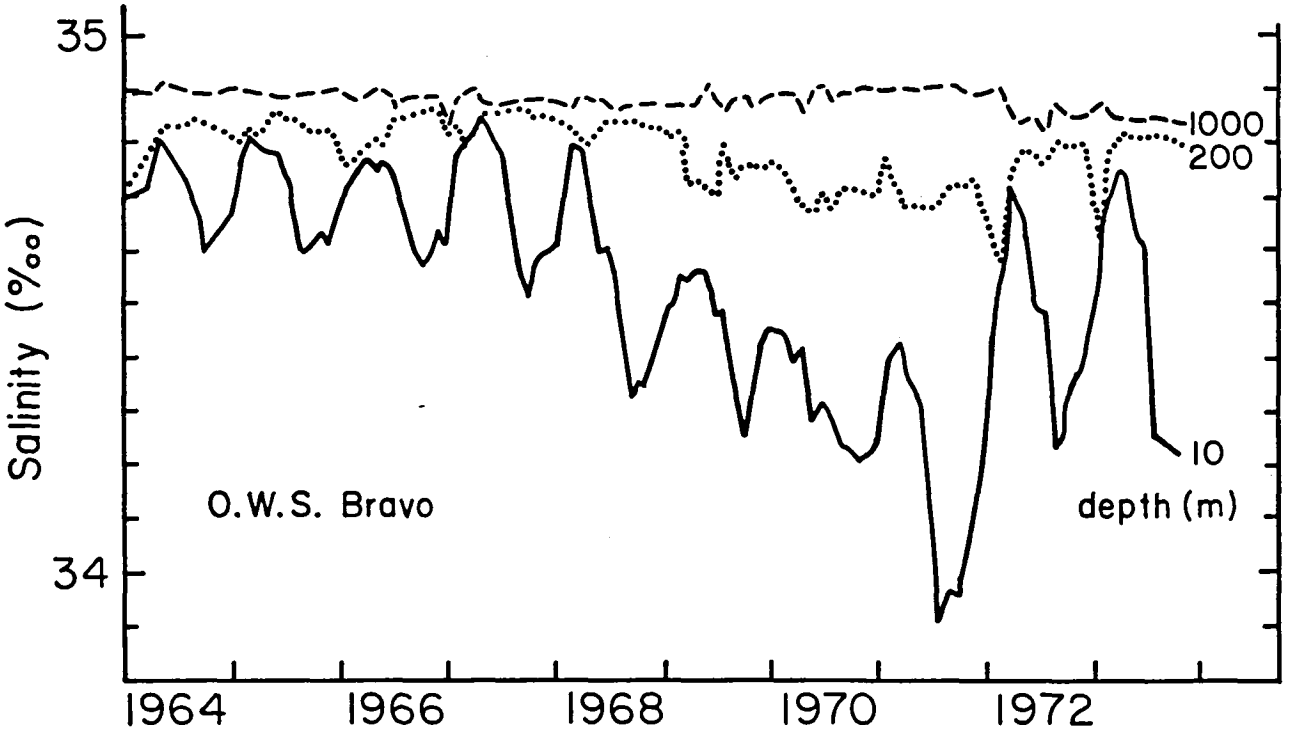

Fig. 4. Salinity on various levels as measured at OWS Bravo (after Lazier, 1980).

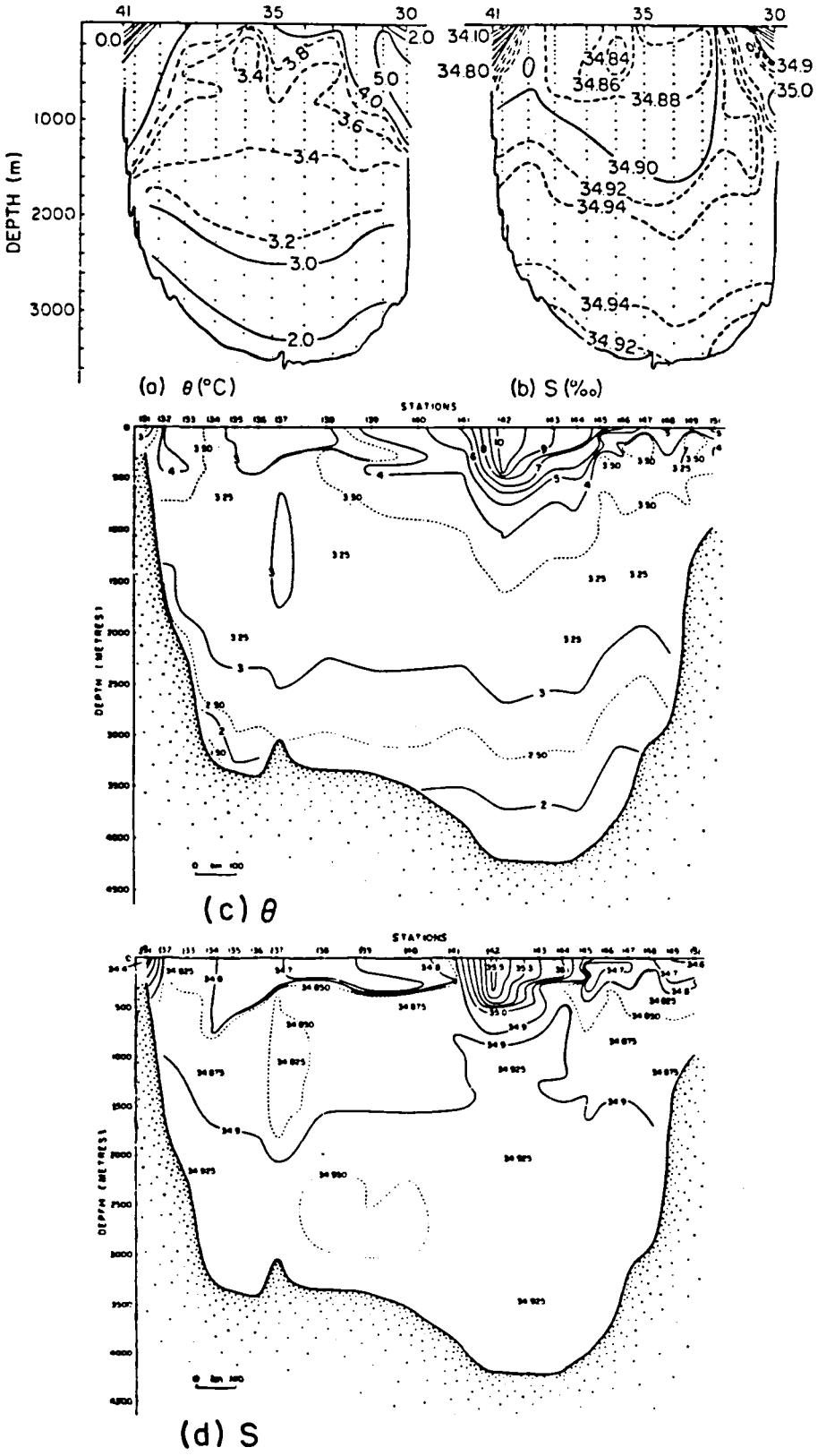

Fig. 5. (a) Potential temperature, (b) Salinity in April, 1966, (c) Potential temperature, and (d) Salinity in April, 1978 along a section from Cape Farewell to Flemish Cap.

The fact that the Labrador Sea Water mass all cooled by 0.5°C in less than a decade indicates that the intermediate waters of the ocean can respond to climatic changes over periods of a few years. In an analysis of the subsurface data collected between 1964 and 1974 at OWS Bravo, Lazier (1980) showed that beginning in 1967 or 1968, the salinity of the upper few hundred meters of the Central Labrador Sea began to decrease at a rate faster than 0.1°/oo/year at 10 meters (Figure 4). At 1,000 metres, salinity began increasing at a rate of 0.01°/oo/year. In the winter of 1972, the upper layer values returned to their more normal range while 1,000 metre values decreased in salinity to values 0.04-0.06°/oo less than those seen in the period 1964-1966.

Lazier (1980) suggested that during the late 60's, increased Arctic outflow or increased offshore transport of low salinity Arctic outflow from the Labrador and Greenland currents spread a low salinity surface layer over the Central Labrador Sea. This increased the vertical stability within the upper water column and surpressed convection. Over a series of mild winters, the vertical stability increased making it more difficult for convection to take place. Finally, the intense winter of 1972 broke through the stability and mixed the fresh water deep into the water column. This low salinity input during the late 1960's resulted in the low salinity LSW that was observed forming in 1976.

This freshening was not confined to the Labrador Sea. The section from Cape Farewell south to Flemish Cap (Figure 5) shows that the entire upper 2 km of water column except for the warm salty North Atlantic Current has decreased its salinity and cooled. It is also possible that there has been some freshening of the NADW near the boundaries, especially the northern one. This freshening therefore affected the Irminger Sea Water formed to the east of Greenland as well as the Labrador Sea Water and perhaps the waters formed north of Iceland and overflowing into the North Atlantic. What we are seeing is not just one component of the northern intermediate water masses changing; we are seeing all of them changing and this implies a change imposed over a large area of the sub polar Atlantic Ocean.

CHEMICAL SIGNATURES ASSOCIATED WITH THE FRESHENING OF NORTHERN ATLANTIC WATERS BETWEEN 1972 AND 1981

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Further insight into the possible causes of the freshening of deep water in the northern Atlantic between the GEOSECS survey (1972) and the TTO survey (1981) is provided by measurements of dissolved silica and dissolved oxygen. As all the northern end member waters are low in dissolved silica content and high in dissolved oxygen content, the message carried by these chemical properties of silica has to do with the extent to which the waters in the northern Atlantic have mixed with the high silica content waters of Antarctic origin. A change in the degree and pattern of this intrusion is noted between the GEOSECS and TTO surveys.

On the map in Figure 1 the locations of GEOSECS stations 8, 5, 4, 2, 27 and 24 are superimposed on the TTO track map. In Figure 2 through 7 comparisons are made of the salinity, initial phosphate ($PO_4°$), dissolved silica (H_4SiO_4) and apparent oxygen utilization (AOU) values for these GEOSECS stations and the nearest TTO stations. In each case potential temperature serves as the reference parameter.

Along a traverse southward from the Denmark Straits the following relationships are seen. At 60°N, 56°N, and 54°N (Fig. 2), the TTO salinities are distinctly lower than the GEOSECS values but no significant differences in the chemical signatures are found. At 47°N (Fig. 3, upper panel) both a salinity difference and dissolved silica and oxygen differences are seen between the surveys. The TTO waters are lower in silica and in oxygen (i.e. they have a smaller component of Antarctic derived water) than the GEOSECS waters. At 42°N (Fig. 3, middle panel) the GEOSECS-TTO salinity difference is gone. However, the TTO stations have less silica and higher oxygen. In Figure 4 contours of H_4SiO_4 content along the 45.86°/oo and 45.83°/oo isopycnal horizons are shown based on the TTO data set. Also shown are the H_4SiO_4 values obtained for these densities at the GEOSECS stations. As can be seen, the GEOSECS values for station 1 (45°N) are considerably higher (i.e., ≈ μm/kg) than expected from the TTO contours. The results along this traverse can be best explained by calling upon intensified overflow of water across the Denmark Straits into the western basin. This intensification results in a push back of the intruding high silica-low oxygen southern water.

The situation at station 24 in the Gibbs Fracture Zone is quite different. The relatively warm, saline and low silica content water found in the bottom of the fracture zone during the GEOSECS survey has been displaced by relatively cold, fresh and high silica content water. This water can only have originated

in the open Atlantic to the south of the fracture zone. Our interpretation is that as the pulse of new water from the Denmark Straits pushed southward along the western margin of the basin, old water moved northward along the eastern margin of the basin and entered the western end of the Gibbs Fracture Zone. It appears that the intensification of production of deep water at the head of the western basin was accompanied by the weakening of the production of deep water at the head of the eastern basin!

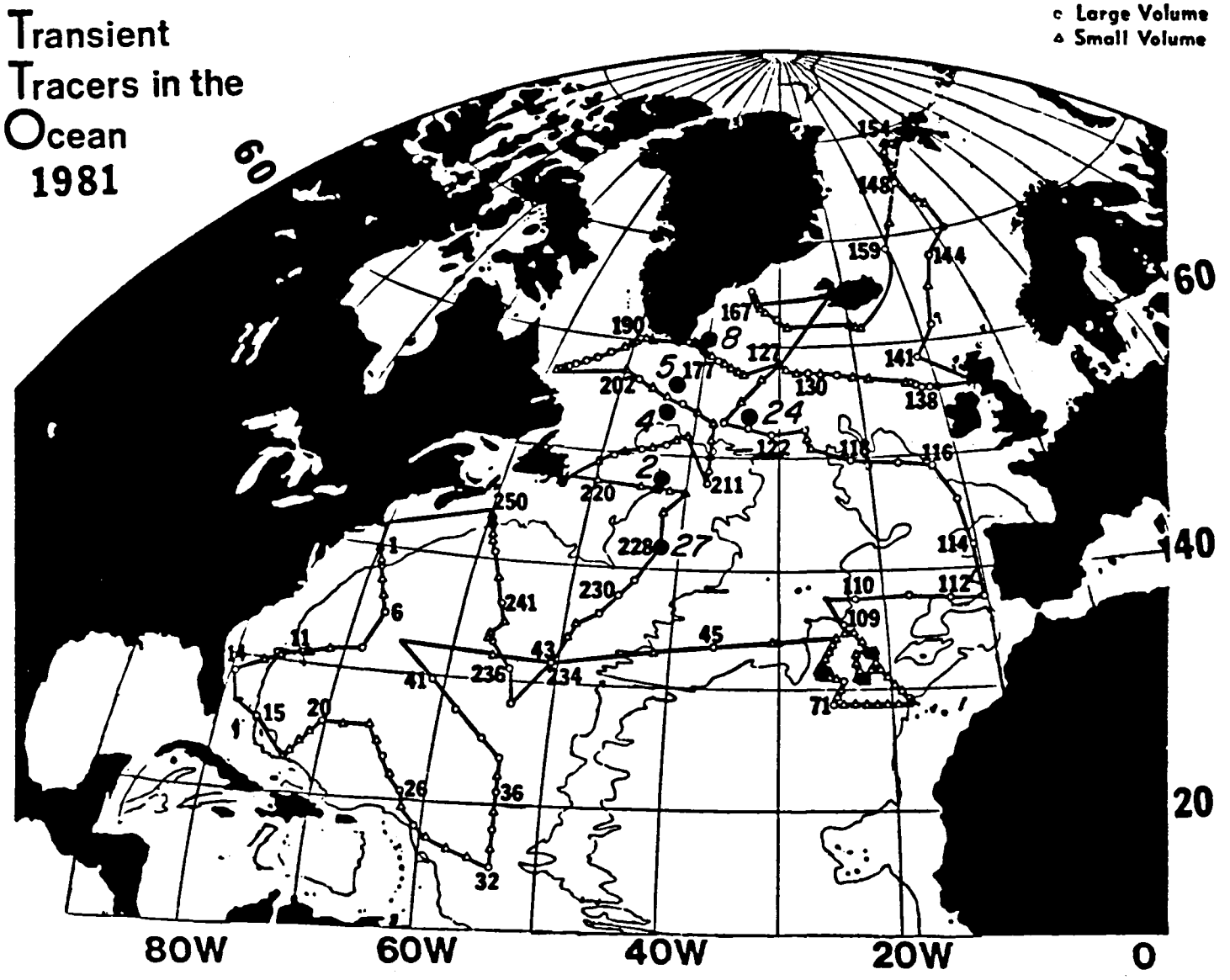

Fig. 1. Locations of GEOSECS stations 8, 5, 4, 2, 27 and 24 superimposed on the TTO track.

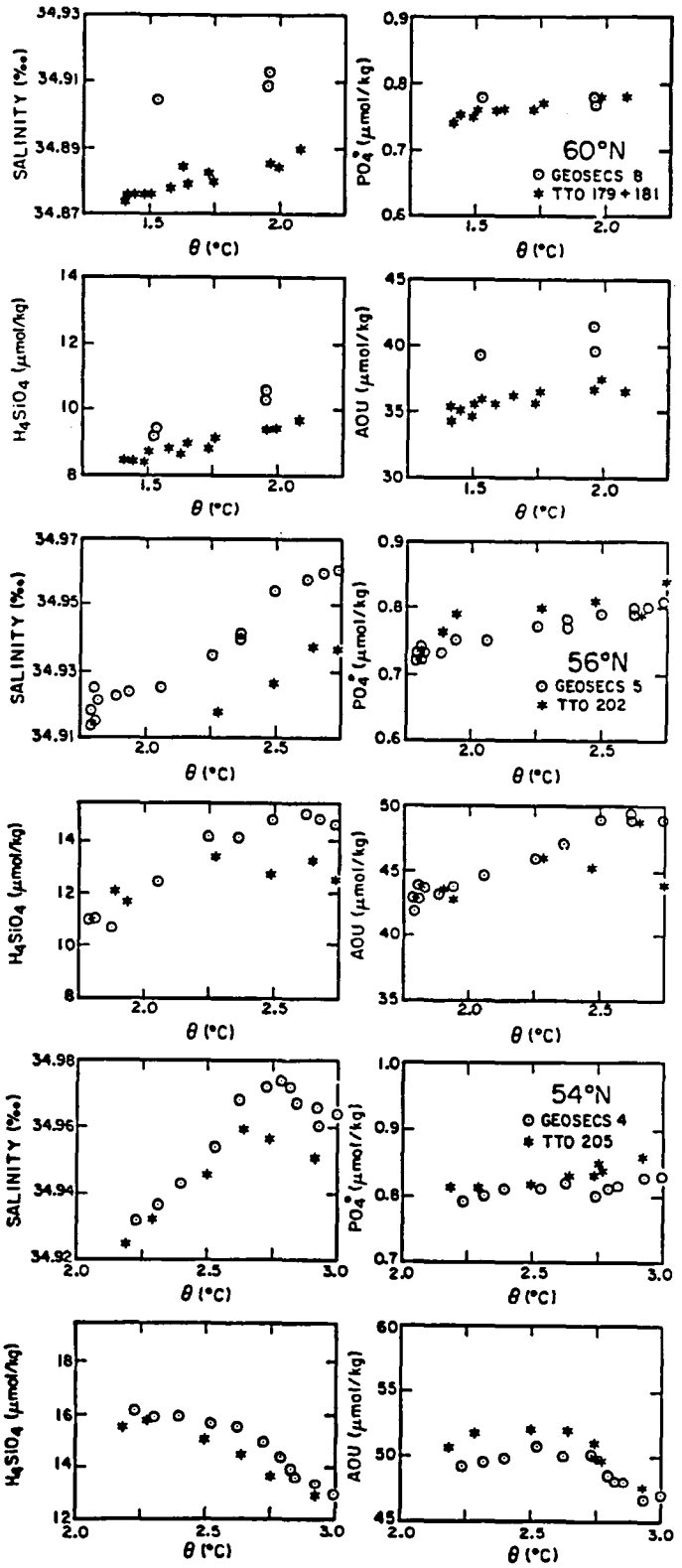

Fig. 2. Salinity, initial phosphate ($PO_4°$), dissolved silica (H_4SiO_4) and apparent oxygen utilization (AOU) for three GEOSECS stations and nearest TTO stations.

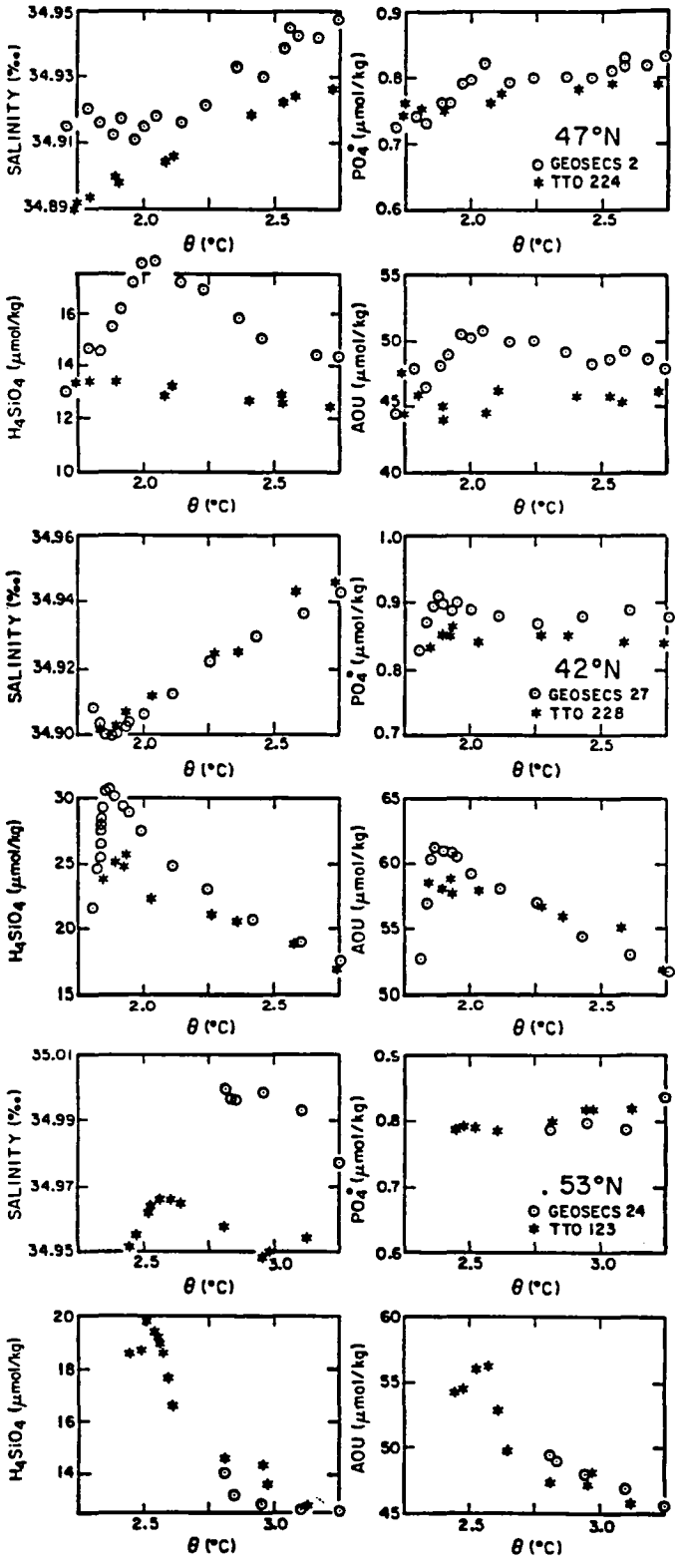

Fig. 3. Same as Fig. 2 for three additional locations.

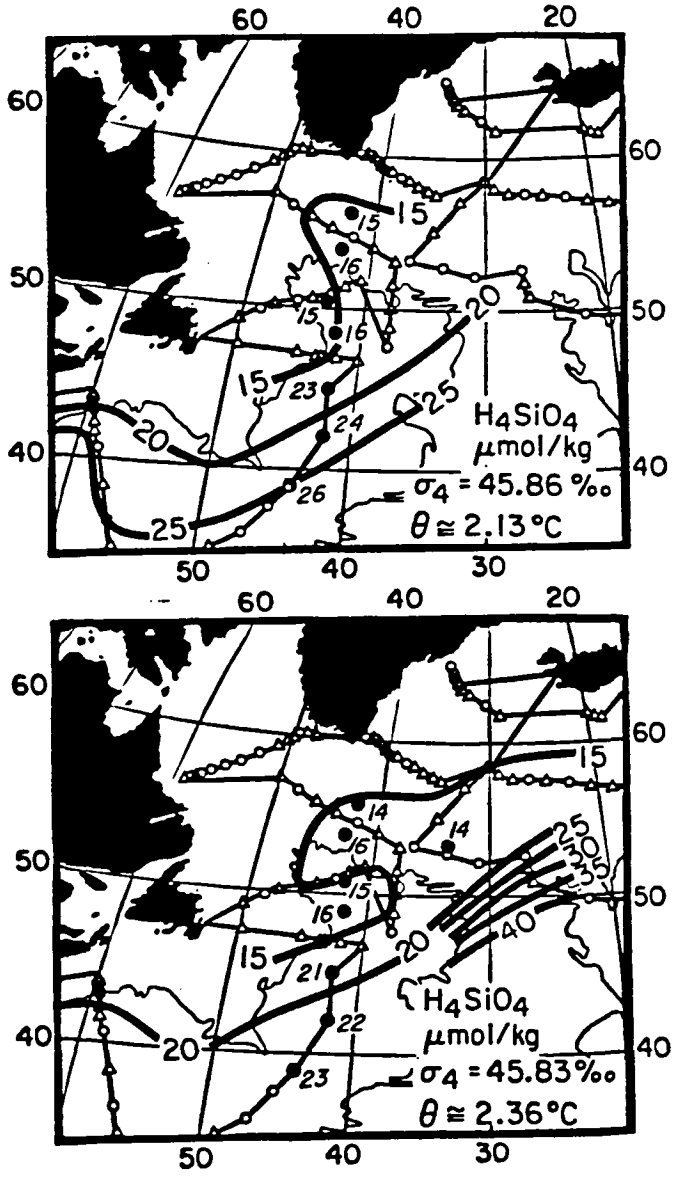

Fig. 4. Contours of H_4SiO_4 along 45.86°/oo and 45.83°/oo isopycnal horizons based on TTO data and corresponding values at the GEOSECS stations.

NORTH ATLANTIC DEEP WATER AND THE WORLD OCEAN

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North Atlantic Deep Water (NADW) by being warmer and more saline than the average abyssal water parcel introduces heat and salt into the abyssal ocean. The source of these properties is upper layer or thermocline water considered to occupy the ocean less dense than sigma-θ of 27.6 (Gordon and Piola, 1983: Fig. 1). That NADW convects even though it's warmer than the abyssal ocean is obviously due to the high salinity. In this way, NADW formation may be viewed as haline convection. The counter "force" removing heat and salinity (or introducing fresh water) is usually considered to take place in the Southern Ocean where upwelling deep water is converted to cold fresher Antarctic water masses. The Southern ocean convective process is driven by low temperatures and hence may be considered as thermal convection. A significant fresh water source may also occur in the North Pacific where the northward flowing of abyssal water from the Southern circumpolar belt is saltier and denser than the southward flowing, return abyssal water. The source of the low salinity input may be vertical mixing of the low salinity surface water or the low salinity Intermediate water.

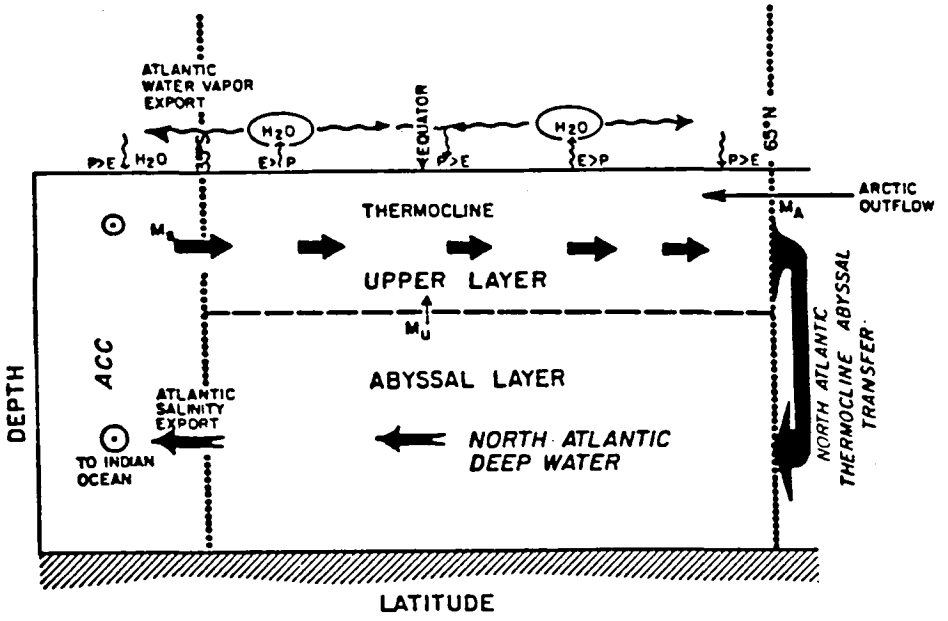

Fig. 1. Schematic diagram of the meridional circulation model for the Atlantic Ocean. The upper layer is required to flow northward from 35°S across the equator and into the northern North Atlantic in order to conserve the mass lost by formation of North Atlantic Deep Water. The salinity of the upper layer is altered by freshwater exchange with the atmosphere. The Atlantic abyssal layer exports the relatively salty NADW into the Antarctic Circumpolar Current and atmospheric water vapor is exported southward across the 35°S and eventually contributes in decreasing salinity of the upper layer south of Africa. The parameters M_S, M_U, M_A are described in the text.

It is likely that heat and salt are also introduced into abyssal water by vertical mixing, particularly for these thermocline regions susceptible to salt finger instability. The Intermediate water mass of low salinity is a participating factor. If NADW were turned off and another convective transfer of water from upper to abyssal layer is not established, it is reasonable to conclude that the thermocline water would become saltier (assuming the net atmosphere fresh water forcing remains unchanged). In this situation further enhancement of salt finger activity which drives large K_z values would increase the transfer of thermocline heat and salt into abyssal water, compensating to some measure for the absence of NADW. In this case the abyssal ocean may not be altered drastically in regard to heat and salinity, though some changes in the non-conservative properties may occur, since the transfer mechanism switches from point source convection to broad regional diffusion.

In today's ocean simple balances suggest that the primary transfer of heat and salt into the abyssal ocean is not necessarily NADW but rather cross thermocline mixing (Gordon, 1975).

The other aspect of NADW process which hasn't received much attention is the return route of abyssal water to the upper layer and back to the North Atlantic. The standard picture presented is one in which abyssal water upwells, migrates to the Antarctic Circumpolar Current (presumably the sub-Antarctic zone) for advection to the Atlantic equator (which from energy considerations would mainly be in the lower thermocline or Antarctic Intermediate Water). This is essentially a cold water route. An alternative warm water route may be effective, as brought to light by recent global fresh water budgets for the ocean (Piola and Gordon, 1984) and field work in the region of the Agulhas retroflection south of Africa. By this route abyssal water upwells into the thermocline at a rate to balance NADW formation rate. It is warmed and generally made saltier. Pacific to Indian transfer of thermocline water occurs in the Indonesian Passages (Fig. 2) and transfer into the Atlantic is achieved by an arm of the Agulhas current which does not participate in the general retroflection circulation pattern (Fig. 3).

The warm water route has significant impact in inter-ocean heat and fresh water transfer. An effective role which the ocean may play in global climate is the ability to transfer these properties across longitudes on a hemispheric scale. The warm water route essentially moves heat and low salinity water from the Pacific to the Atlantic by way of the Indian Ocean.

The large oceanic to atmosphere heat transfer in the North Atlantic is derived from the Pacific Ocean, and the net evaporation of the Atlantic is balanced by the net precipitation of the Pacific.

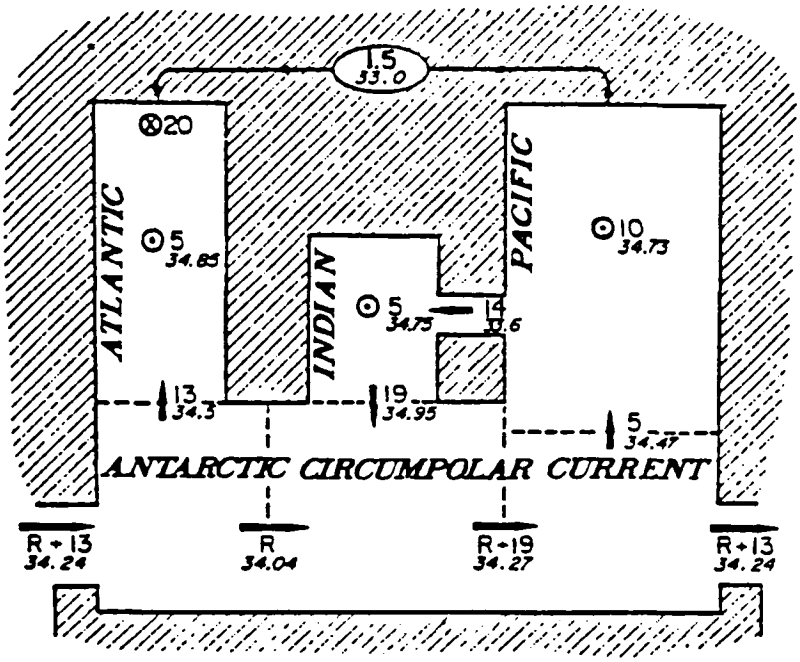

Fig. 2. Schematic diagram of the upper-layer ocean circulation based on the mass and fresh-water balance of each ocean north of the Antarctic Circumpolar Current. Upper layer transports and salinities are given through the southern boundary of each ocean and through the 27.6σ surface which divides the upper layer from the deep waters. R represents the recirculation in the upper layers within the Antarctic Circumpolar Current.

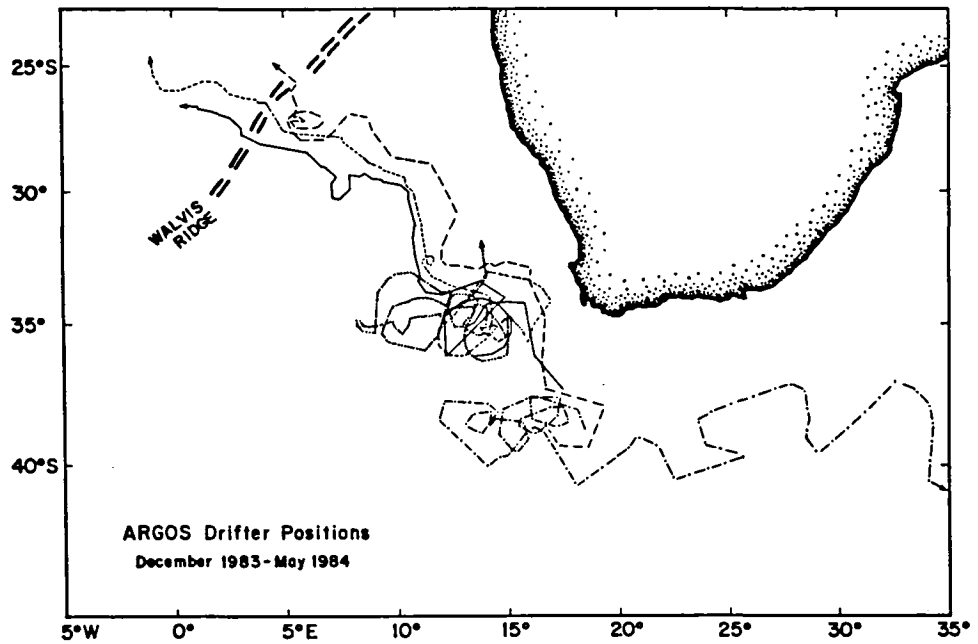

Fig. 3. Trajectories of satellite tracked drifters set out in November 1983 within the Agulhas retroflection system. Three drifters set within Indian Ocean water moved at 0.5 to 1 Kt into the South Atlantic. The hydrographic data also shows a large flux of warm Indian Ocean into the South Atlantic. Does this situation represent the mean or only an occasional event, and does the Indian Ocean inflow compensate for sinking in the North Atlantic.

In order to evaluate the impact of a change in NADW to global climate it is necessary to:

1. Determine what percentage of heat and salt input to abyssal ocean is accomplished by the NADW process.

2. What is the "thermohaline" path followed by the return flow of upper layer water to the North Atlantic.

OCEANOGRAPHIC VARIATIONS IN RECENT DECADES AND THEIR IMPACT ON THE FERTILITY OF THE ICELAND SEA

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Due to its location at or near the oceanic polar front, the Iceland Sea (Fig. 1) is an area particularly sensitive to climatic changes. In warm years a strong influx of Atlantic water from the Irminger Sea can be traced all along the North Icelandic shelf area. This inflow is mainly determined by past meteorological conditions at the west and north coasts of Ireland (Steffánsson, 1962; Stefánsson and Gudmundsson, 1969). In periods when the ice belt along the east coast of Greenland is relatively narrow, the surface layers between Iceland and Jan Mayen consist of arctic water with practically no polar component. Conversely, in cold periods an appreciable proportion of cold, low-salinity polar water is found in this area, and large parts of the North Icelandic shelf area may be covered with drift ice. Sea surface temperatures north of Iceland, especially in spring, are closely correlated to the frequency and extension of drift ice (Stefánsson, 1969; Malmberg, 1972).

Changes during this century in North Icelandic waters have been characterized by i) a general warming up during the first decades (Stefánsson, 1969) culminating in the thirties; ii) a slight downward trend in the forties and fifties, followed by a greatly increased extension of drift ice and a marked lowering of temperature and salinity in the period 1965-1971; and iii) in the last decade very cold years have alternated with relatively mild ones (Figs. 2-3). These variations can be traced down to a depth of 100-200 m, both in the North Icelandic shelf area and in the deep area between Iceland and Jan Mayen (Stefánsson, 1969; Stefánsson, 1968). At 200-500 m in that region the temperatures have in general been slightly higher since 1965 than in the preceding period, presumably as the result of shallower winter convection due to lower salinities in the surface layers. There is no significant difference between the mean salinity (34.92) at 1000 m in the two periods 1949/1957 and 1973/1980.

Since the overflow across the Iceland-Greenland Ridge probably consists largely of Arctic intermediate water including winter water from the Iceland Sea (Stefánsson, 1968; Mann, 1969; Swift et al., 1980), changes in the salinity of that water may be reflected in the Overflow water. On the other hand, very rapid changes are known to exist in the Overflow, both with regard to T,S-characteristics and velocity. Hence the importance of monitoring both the short-period and long-period variations of the Overflow should be emphasized. At the same time the causes of short-periodic fluctuations need to be studied.

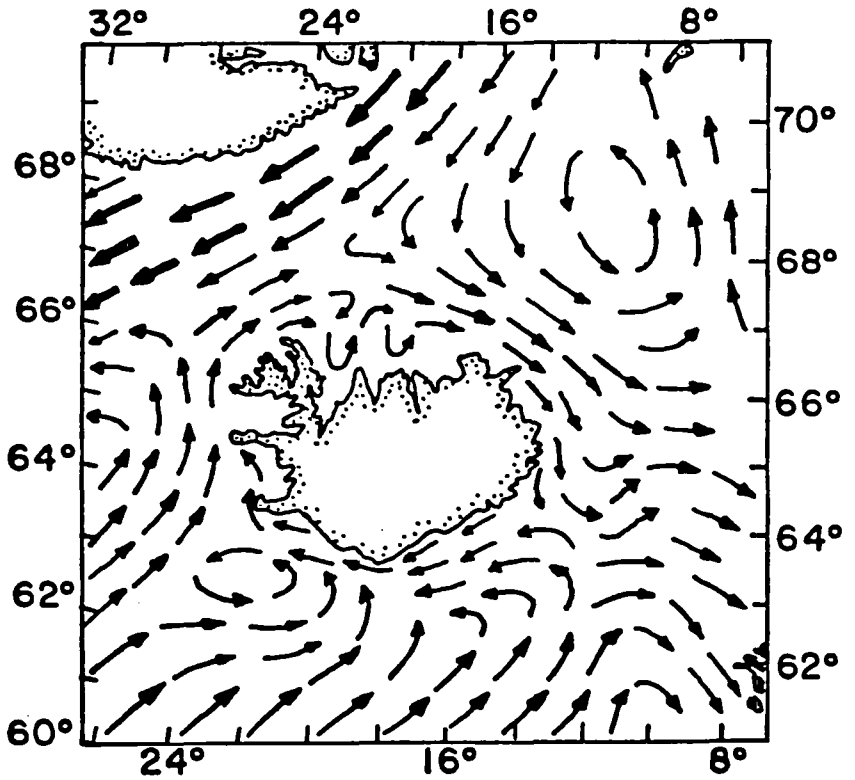

Fig. 1. The average residual surface currents of the Iceland Sea referred to the warm period before 1965 (Stefánsson, 1961).

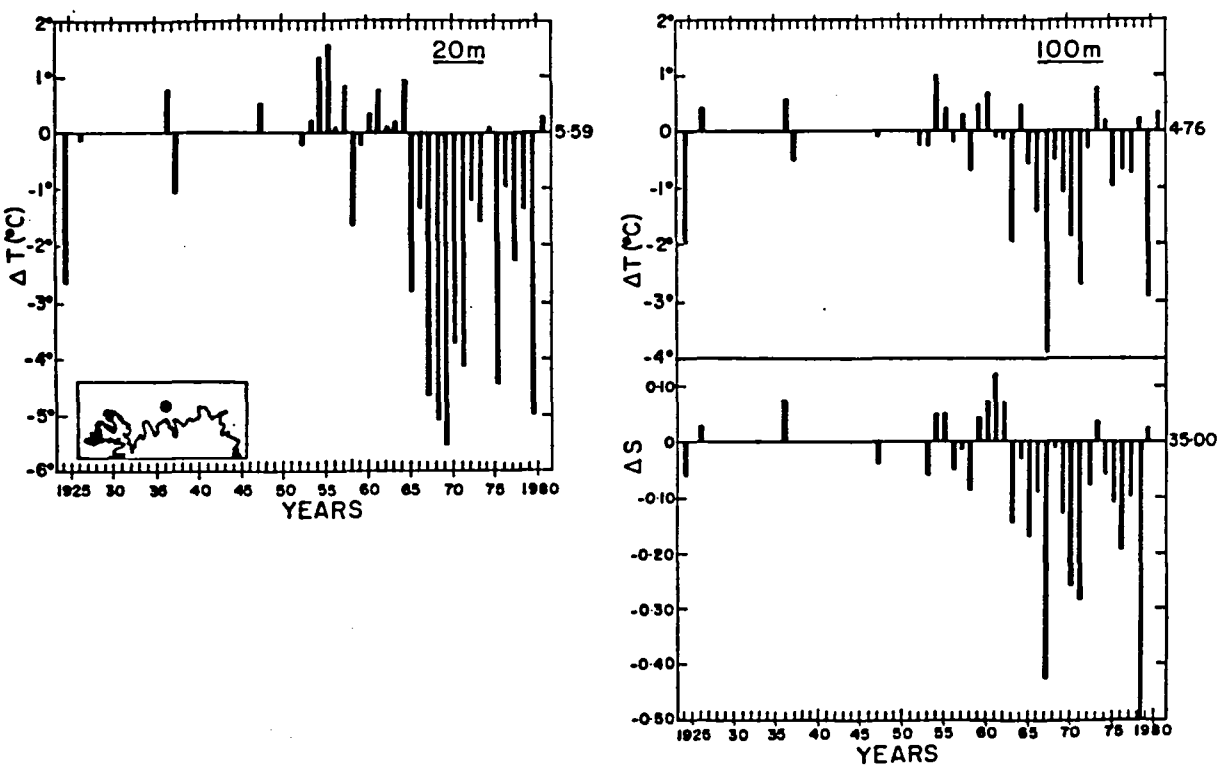

Fig. 2. Temperature (20 and 100 m) and salinity (100 m) variations at S-3 (66°32'N, 18°50'W) in June in the period 1924-1980 referred to the 1924-1960 mean.

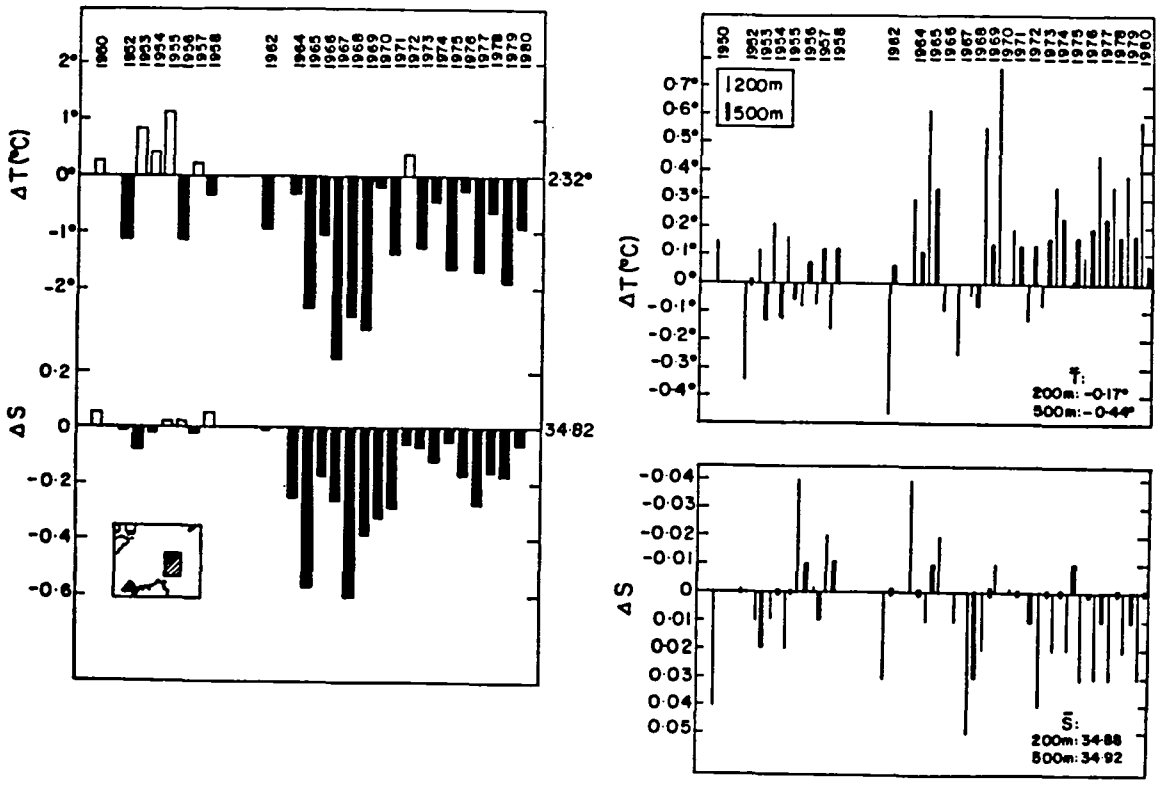

Fig. 3. Temperature and salinity anomalies in June within the rectangular area between 67° and 69°N, and 11° and 15°W referred to the 1950-1958 mean. Left: Variations 25 m; right: variations at 200 and 500 m. After Malmberg (1983).

Climatic fluctuations affect life in the sea both directly and indirectly. In particular, it seems that the changes of recent decades in the density stratification of the surface layers have had marked biological implications. This has been confirmed by studies of primary production over the last 25 years in North Icelandic waters (Thordardottir, 1980). For the period late May - early June very marked year to year differences were found, normally in such a way that the productivity was low in cold years, but high in warm years. This, however, had less to do directly with temperature than with the density stratification of the surface layers (Fig. 4). There is in general a fair degree of correlation between primary production, vertical mixing and nutrient concentrations in the surface layers (Fig. 5). Thus it appears that low primary production occurring in the cold years can be attributed to marked stratification, preventing the renewal of nutrients in the surface layers, whereby continued growth of phytoplankton was greatly reduced. There are, however, certain discrepancies which need to be examined. Thus small nutrient concentrations do not necessarily imply very strong stratification. During periods of intense productivity nutrients may simply be maintained at a low level, even though they

are to a moderate degree diffusing upwards through a weak pycnocline. The effect of grazing will also complicate the simplified relationships. Thus the primary production was on the average considerably smaller during the warm years of the period 1958-1964 than during those years which were warm in the period 1970-1980. It has been postulated (Thordardottir, 1977) that this was probably due to the effect of zooplankton grazing, since zooplankton densities were much greater in the former period than in the latter, when e.g. Calanus finmarchicus which constitutes the main food of herring practically disappeared from the area. Changes in the composition of the phytoplankton could possibly account for the low densities of zooplankton in recent years (Ástthórsson et al., 1983), but further studies on these relationships are under way. Finally, the changes in food conditions as the result of decreased primary production were probably the main cause of the collapse of the summer herring fisheries (Jakobsson, 1980).

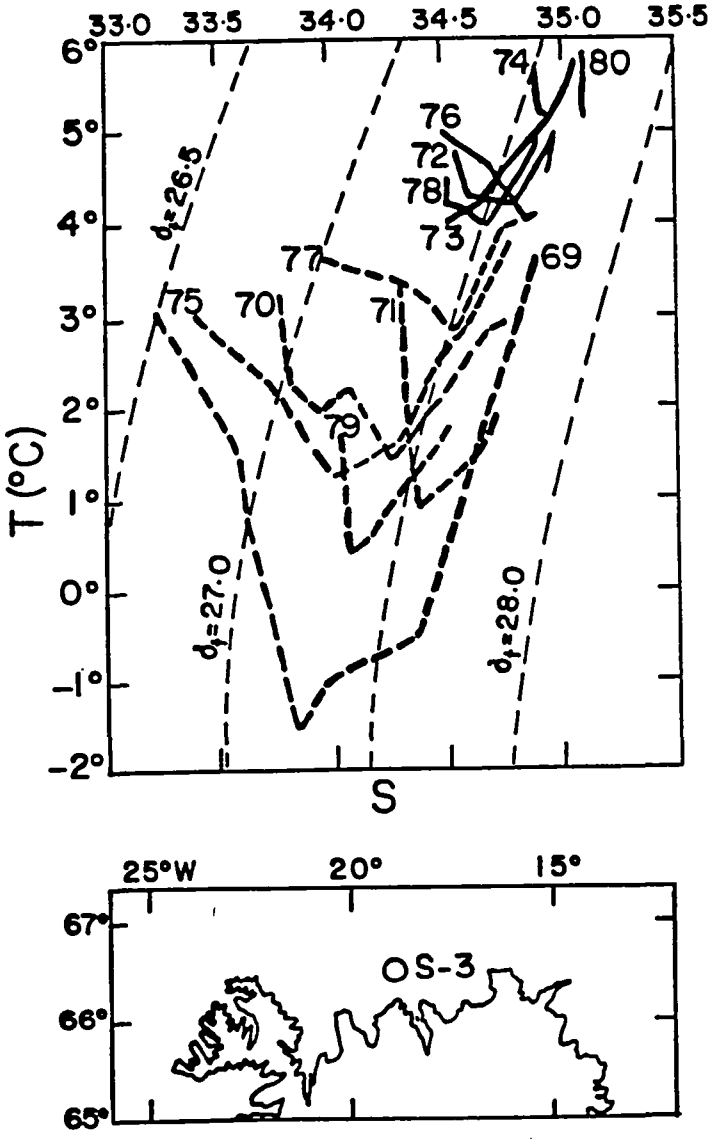

Fig. 4. Temperature-salinity relationships at S-3 (66°32'N, 18°50'W) in May/June for different years and periods. Solid curves: warm years; broken curves: cold years. Based on Thordardottir (1980).

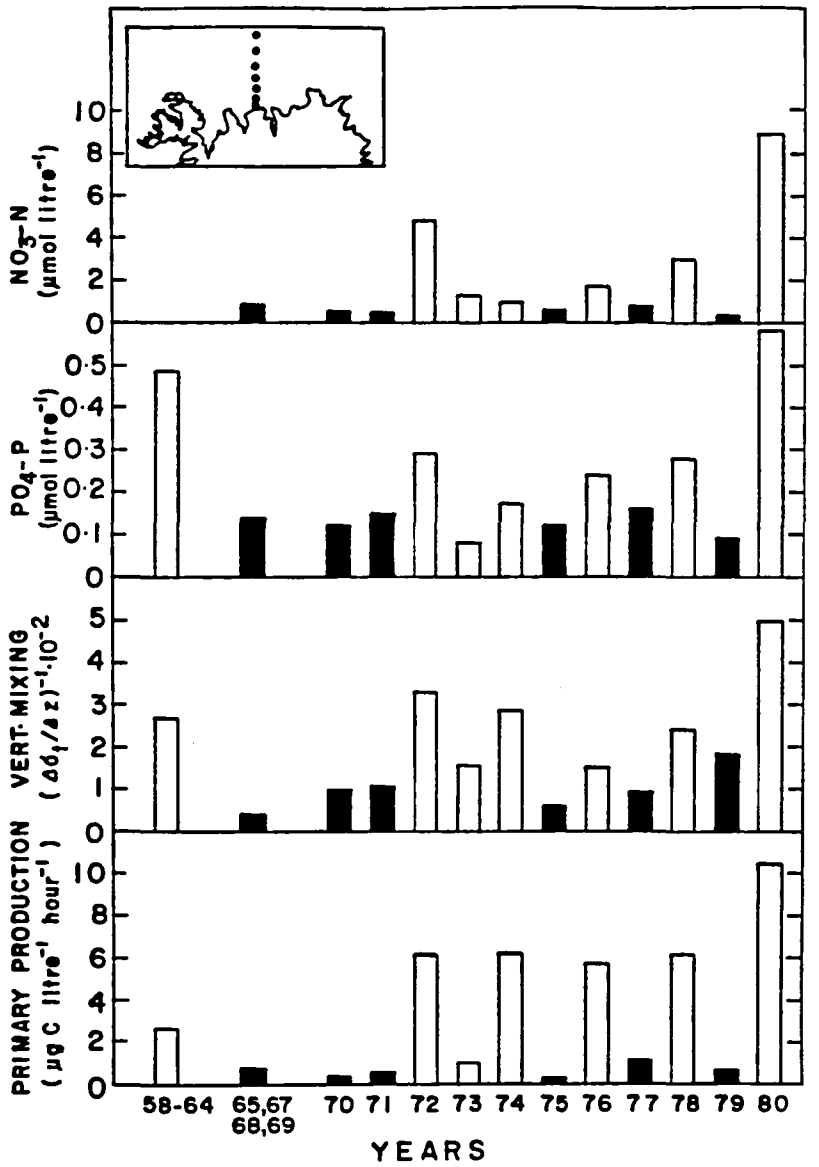

Fig. 5. Comparison between mean primary production, index of vertical mixing (the reciprocal of vertical stability), concentration of nitrate and concentration of phosphate, in May/June for the section off Siglunes (mean for 7 stations). White columns: warm years; black columns: cold years. Based on Thordardóttir (1980).

It is concluded that climatic changes may not only lead to marked changes in the physical and chemical environment of arctic regions, but also have great biological and economical consequences.

NORTH ATLANTIC DEEP WATER FORMATION: INFORMATION FROM ICE CORES

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In the following, the main results of measurements of the CO_2 concentration of air occluded in natural ice during periods of climatic change are presented, as well as other measured ice core parameters. Elements of an interpretation of the data in terms of mechanisms of changing environmental systems are briefly discussed.

I. CO_2 concentration measurements on air occluded in ice cores (Greenland and Antarctica)

The last 100 y

- Preindustrial value (1800-1850): 280±5 ppm (measurements on samples of different size with laser IR-spectroscopy and gas-chromatography).

- During last 1000 y fluctuations are not to be excluded, but are probably less than 10 (to 20) ppm (averaged over the 20 y required for gas enclosure).

The glacial-postglacial transition (Fig. 1)

- Low CO_2 concentrations between 25,000 and 15,000 BP: 180 to 200 ppm.

- Transition to values in the 280 to 300 ppm range ca. 13,000 BP, probably almost in phase with $\delta^{18}O$ shift; indications of lower values during period 10,000 to 11,000 BP (Younger Dryas); final transition to Holocene values of 300±20 ppm at around 10,000 BP (Holocene values influenced by higher CO_2 content of melt layer ice).

Rapid climatic variations in the Wisconsin (Fig. 2)

In the period 30,000 to 40,000 BP (approx.) rapid climatic variations reflected in $\delta^{18}O$ are in phase with CO_2 variations: cold periods with 180-200 ppm, warm periods with 250-270 ppm. (Artifact due to CO_2 enriched melt-layers, exchange with carbonates in ice matrix, etc. cannot completely be excluded; final proof should be obtained from analysis of Antarctic ice cores with different chemical properties and accumulation temperatures.)

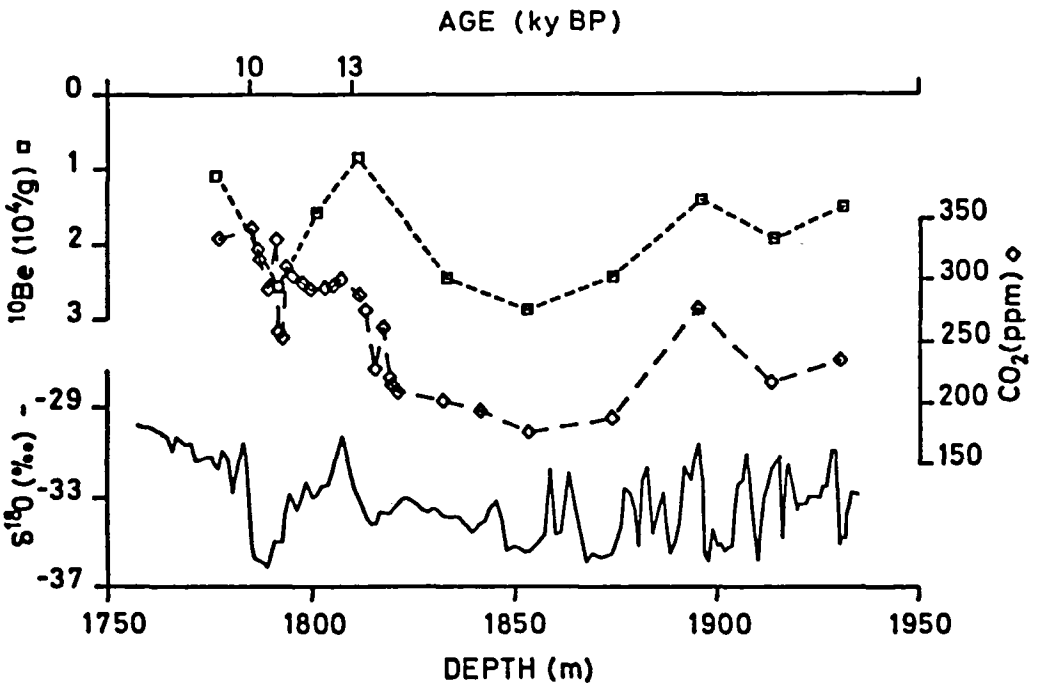

Fig. 1. ^{10}Be concentrations (10^4 atoms per g of ice), CO_2 concentrations (ppm) and $\delta^{18}O$ as measured in the Dye 3 ice core. Top: Tentative time marks, as suggested by the comparison with European lake sediments (Fig. 4).

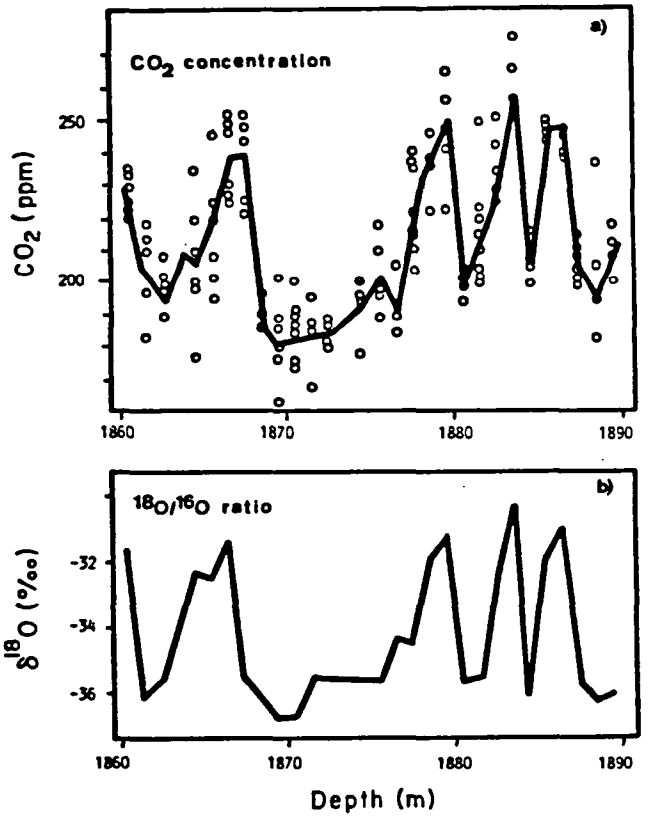

Fig. 2. CO_2 and $\delta^{18}O$ values measured on ice samples from Dye 3: a) Circles indicate the results of single measurements of the CO_2 concentration of air extracted from ice samples. The solid line connects the mean values for each depth; b) The solid line connects the $\delta^{18}O$ measurements done on 0.1 m core increments.

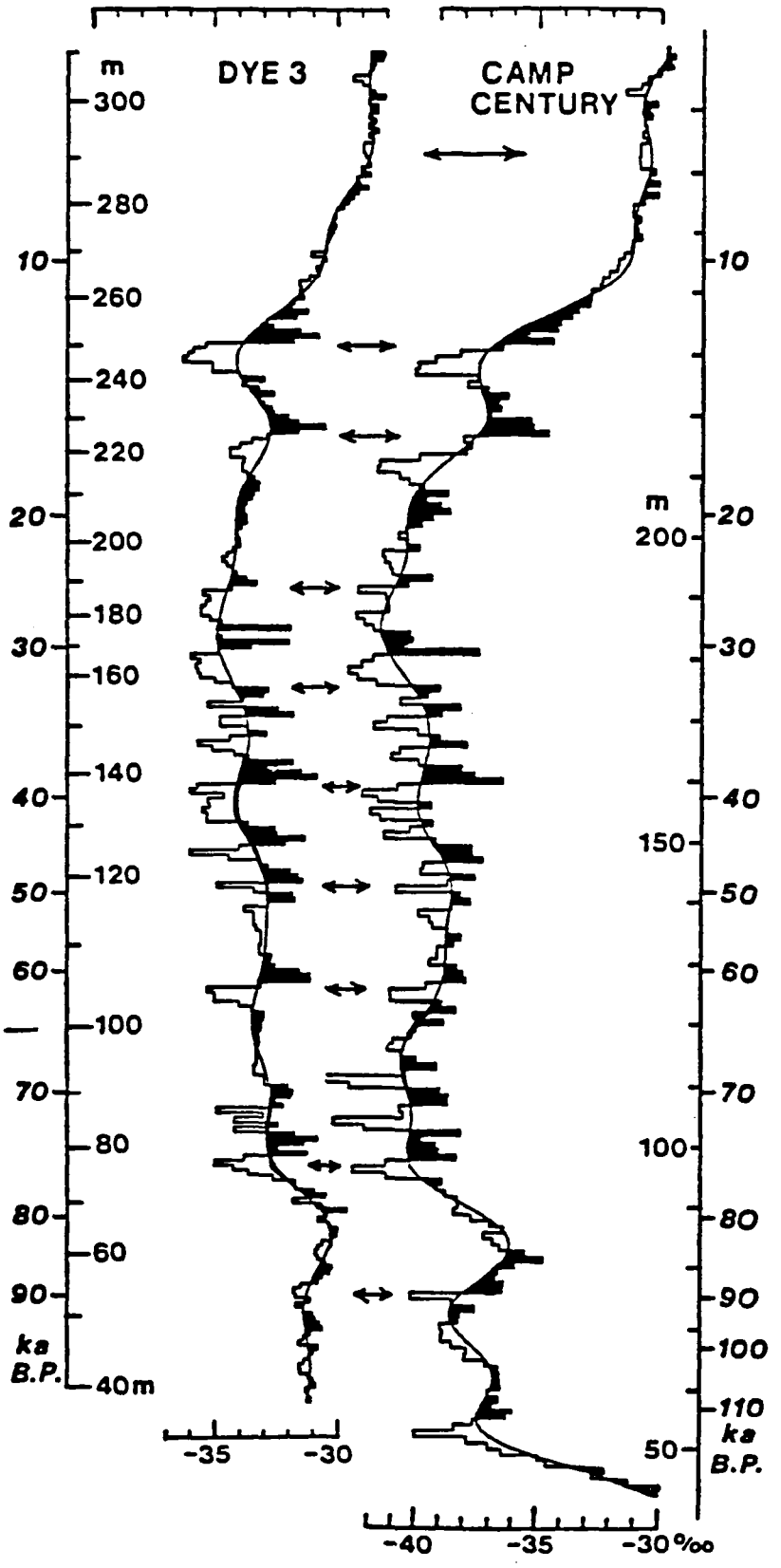

Fig. 3. Comparison of the deepest few hundred meters of two $\delta^{18}O$ profiles through the ice sheet at Dye 3, SE Greenland and at Camp Century, NW Greenland.

II. Other ice cores parameters during periods of drastic climatic change (Langway et al., 1984 and Dansgaard et al., 1984)

General picture: all measured parameters so far show significantly different values between cold and warm climatic system states:

$\delta^{18}O$ difference: 4-5°/oo, CO_2 concentration enriched by factor 1.4, C SO_4, NO_3 depleted in "warm" state by factors 1.6 to 4, ^{10}Be by factor 1.5 to 2.5, also insoluble dust.

Values in warm and cold system states are in relatively narrow bands compared to difference between the bands. Indication of a bistable climate and environmental system. In European lake sediments (Lake Marl) $\delta^{18}O$ signatures for the 14,000 to 9000 BP period can be found which are highly correlated to the $\delta^{18}O$ signature in the Greenland ice cores (Figs. 3 and 4).

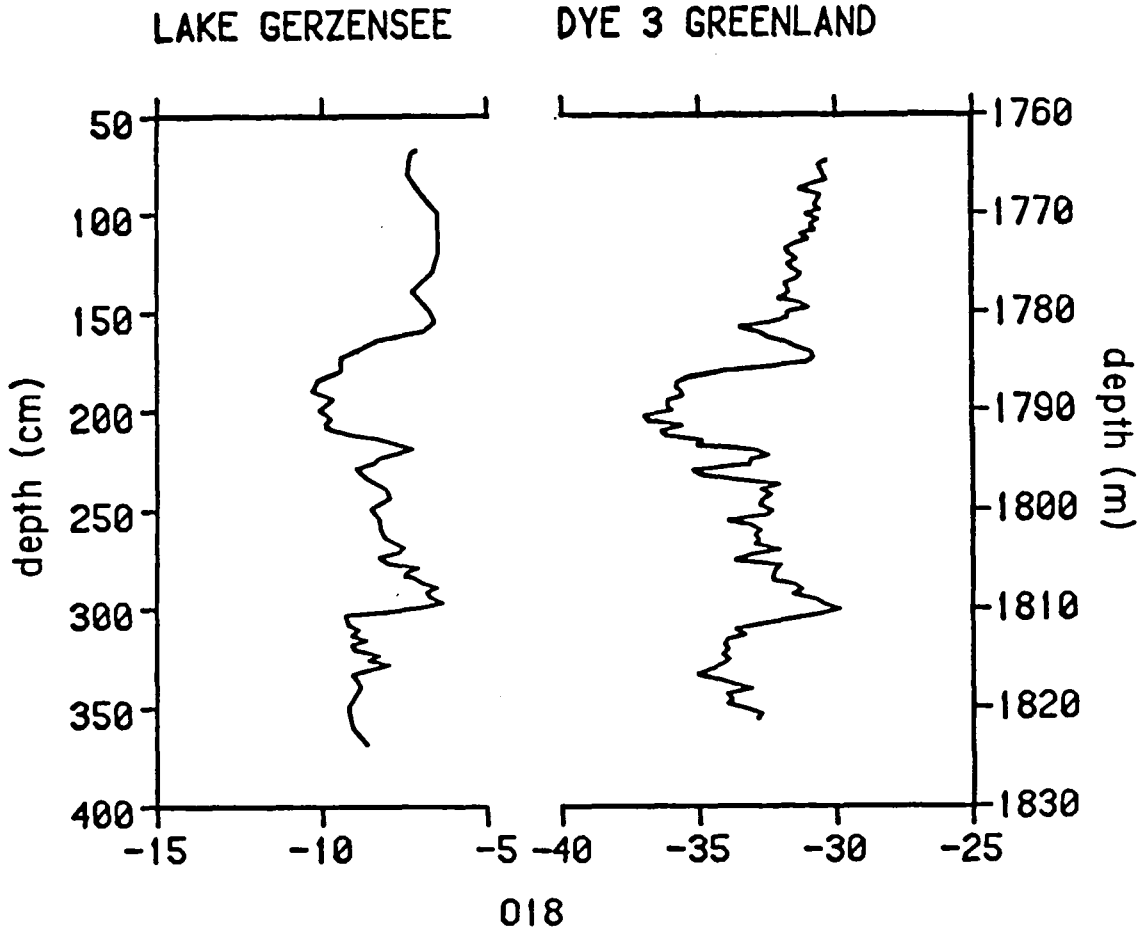

Fig. 4. Comparison of a section of the $\delta^{18}O$ profile from the Dye 3 ice core (right) with the $\delta^{18}O$ record in lake carbonate from Gerzensee (left). The strong similarities suggest that both records represent the same sequence of climatic events and thus the same time period.

III. Elements of explanation of rapid Wisconsin climatic changes

- $\delta^{18}O$ changes (14,000 to 9000 BP) in Greenland ice and European lake sediments with high probability are linked to processes in the North Atlantic Ocean, as deduced from faunal and stratigraphic ocean sediment studies (Ruddiman and McIntyre, 1981), indicating advances and retreats of the deglacial polar water front.

- The earlier rapid changes (reflected in $\delta^{18}O$ profile of Greenland ice cores) might have the same cause, i.e., shifts in the North Atlantic sea ice extension and iceberg influence.

- The system's bistable mode might be the result of lower sea level and higher continental ice mass, influencing ocean circulation in various ways (deep water formation, seabed topography influence).

- "Milankovitch" forcing probably responsible for longer term climatic cycle, e.g., continental ice covers; superimposed undoubtedly are the phenomena of rapid climatic oscillations related to events in the North Atlantic Ocean.

- Deglaciation starting ~15,000 BP, still enabling transitions between cold and warm system states; however, after the last transition around 10,000 BP the system could no longer switch back to the cold state, and the North Atlantic continued its mode of operation till today.

- Changes in North Atlantic circulation might trigger other ocean circulation changes which together might lead to a change in ocean surface chemistry ($\sum CO_2$ and alkalinity) leading to changes of pCO_2 in the ocean surface water and changes of atmospheric CO_2 concentration.

Recommendations

- Importance of paleo-information for understanding of complex environmental system mechanisms is underlined by these ice core, ocean and lake sediment data. They should be continued with high priority.

- Attempts to model isotopic ratios ($^2H/H$ and $^{18}O/^{16}O$ in H_2O) using physical information derived from GCM's. Also modelling of transport of sea salts and continental dust etc. to compare results with ice core information.

- North Atlantic probably crucial area for triggering of general oceanic circulation changes. Study of propagation of North Atlantic changes into the entire oceanic system should not be neglected (equatorial upwelling and mixing of water around Artarctica).

- More realistic modeling of ocean circulation in GCM's, especially regarding heat transport (horizontal).

QUATERNARY NORTH ATLANTIC SURFACE PALEOCEANOGRAPHY IN REGIONS OF POTENTIAL DEEP-WATER FORMATION

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At the time scale of the Quaternary climatic cycles, the sites of formation of North Atlantic Deep Water are not known. The interglacial extreme is presumably exemplified by the modern regions: the Norwegian, Greenland and Labrador Seas. During the major glacial-age coolings in the North Atlantic, the sites may have shifted well to the south, perhaps as far as the limit of the polar front at 40° to 50°N. Still other sites may have been important during intermediate climatic conditions. Because of the close coupling of high-latitude surface waters to North Atlantic Deep Water in the modern ocean (Swift, this volume), the history of sea-surface temperature (SST) oscillations across the high-latitude North Atlantic is relevant to an understanding of deep-water formation on the longer time scales.

The history of the North Atlantic SST changes has been summarized by Ruddiman and McIntyre (1984) and is portrayed schematically in Figure 1. At these time scales, the SST records in all areas are dominated by fluctuations at the orbital periodicities of 100,000 to 41,000 and 23,000 years. North of 55°N, in the regions of modern deep-water formation, extremely cold temperatures persist for very long intervals that are broken only by the minor warmings associated with brief (10,000-year) interglaciations roughly every 100,000 years (Kellogg, 1976). Between 55°N and 45°N, the dominant 100,000-year periodicity is joined by a strong 41,000-year rhythm, and the combined signal from both periods marks the advances and retreats of the polar front. Between 45°N and 35°N, the 41,000-year periodicity disappears, and the still-strong 100,000-year cycle is slightly eclipsed by a remarkably strong signal at the 23,000-year period. South of 35°N, the SST changes are much smaller near the stable center of the subtropical gyre (Crowley, 1981).

Several lines of evidence suggest that these surface signals from the North Atlantic appear in geochemical tracers of Quaternary deep-water change. Mix and Fairbanks (this volume) show that the 41,000-year SST signal prominent in the polar front migrations also appears in the ^{13}C record of benthic foraminifera. Boyle (1984) found a 41,000-year signal in cadmium gradients measured in benthic foraminifera from the South Atlantic. Mix and Fairbanks (in press) also have evidence of the 23,000-year signal in other isotopic tracers. In all cases, these signals are in phase with the North Atlantic SST signals at the comparable regions. The mechanism linking these deep-water responses to the long-term surface-water changes in not yet clear.

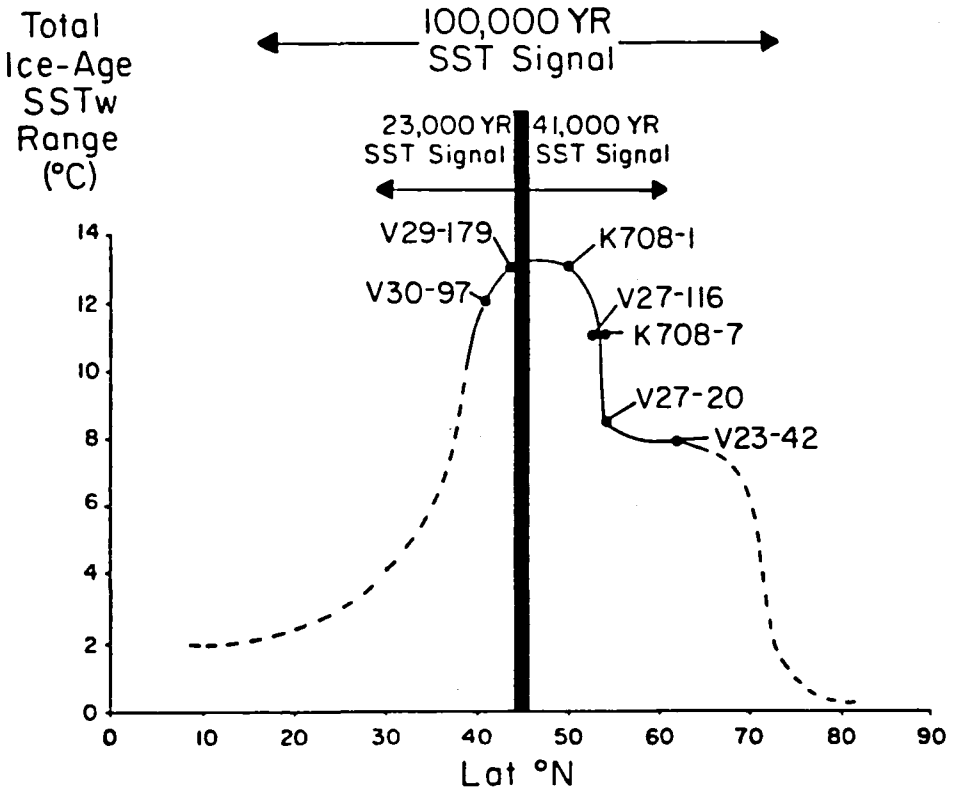
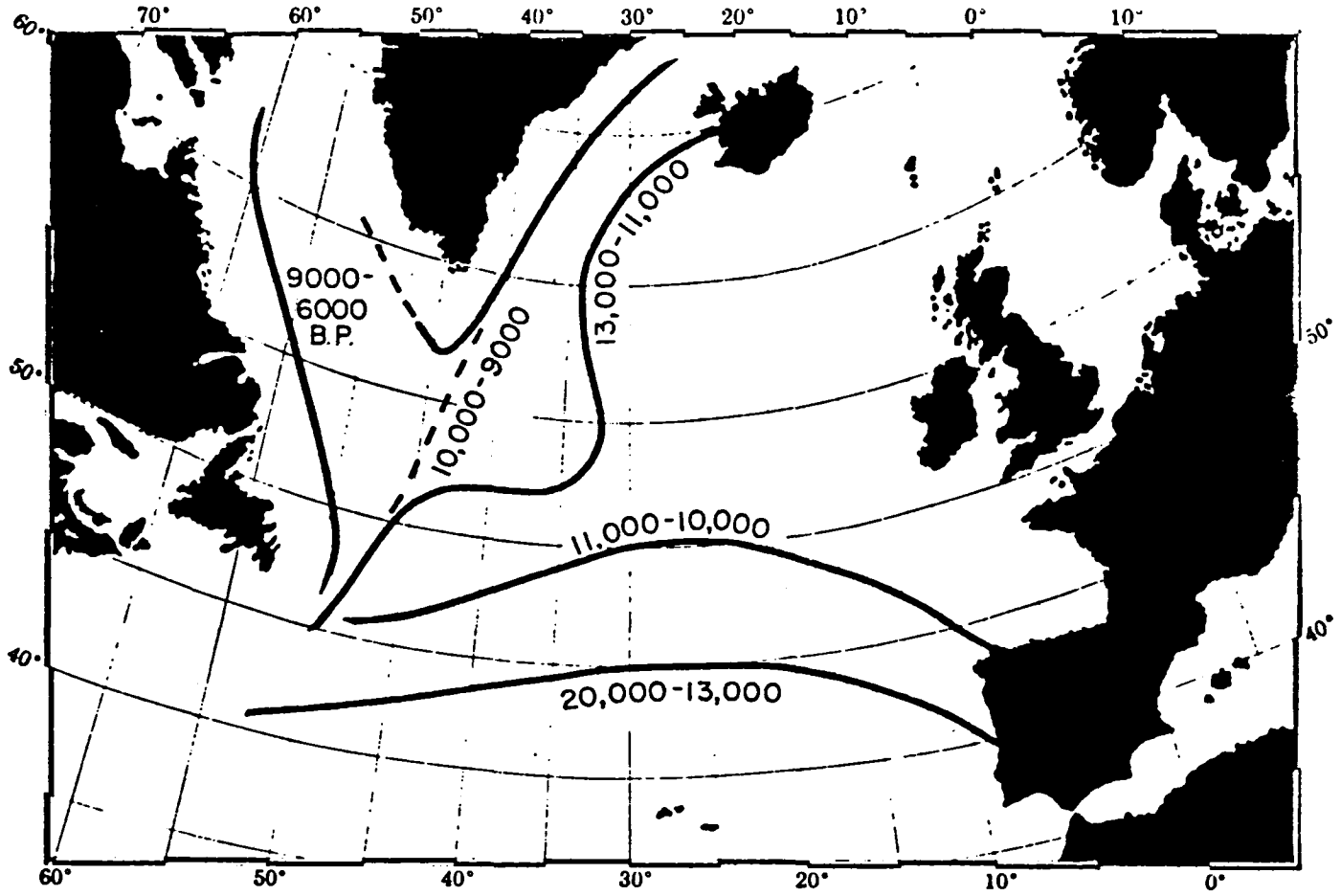


Figure 1. Maximum observed glacial-interglacial difference in estimated summer sea surface temperature versus latitude in the North Atlantic. Dashed portion of curve south of 40°N based on cores reported in Crowley (1981); portion north of 62°N based on cores from Kellogg (1976). Regions of greatest strength of SST changes at the three orbital periods also indicated.

Much shorter fluctuations are also evident in the North Atlantic responses. Ruddiman and McIntyre (1981) documented a cooling lasting less than 1000 years that returned the eastern and central parts of the high-latitude North Atlantic from full-interglacial SST values to a full-glacial condition roughly 10,500 years ago (Fig. 2). This kind of surface-ocean response is also likely to have had a significant impact on the formation of North Atlantic deep water, and it may be more relevant to future changes on a short-term basis.

One critical uncertainty in assessing past changes in deep-water formation is the lack of knowledge about past sea-ice limits. Sea ice is generally reconstructed from negative evidence, such as extreme suppression of productivity of shelled planktonic organisms. Research into more definitive methods of estimating past sea-ice limits is recommended.



DEGLACIAL POLAR FRONT MOVEMENTS

Fig. 2. Map of deglacial retreat and readvances of the North Atlantic polar front (from Ruddiman and McIntyre, 1981). Positions shown were occupied during the major part of the intervals indicated, but transitions between positions were not necessarily instantaneous.

GLACIAL-TO-INTERGLACIAL CHANGES IN NADW FLUXES?

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Interglacial gradients in $\delta^{13}C$ between Atlantic and Pacific deep waters reflect differences between low-nutrient, ^{13}C-enriched North Atlantic Deep Water (NADW) and high-nutrient, ^{13}C-depleted Pacific Deep Water. Reduced Atlantic-Pacific $\delta^{13}C$ and cadmium gradients at the last glacial maximum have been used to suggest substantial replacement of NADW with nutrient-rich Antarctic Bottom Water (Boyle and Keigwin, 1982; Shackleton et al., 1983). We show that the Atlantic $\delta^{13}C$ signal is linked directly to North Atlantic polar-front migration, as reflected by planktonic foraminiferal faunas (Fig. 1).

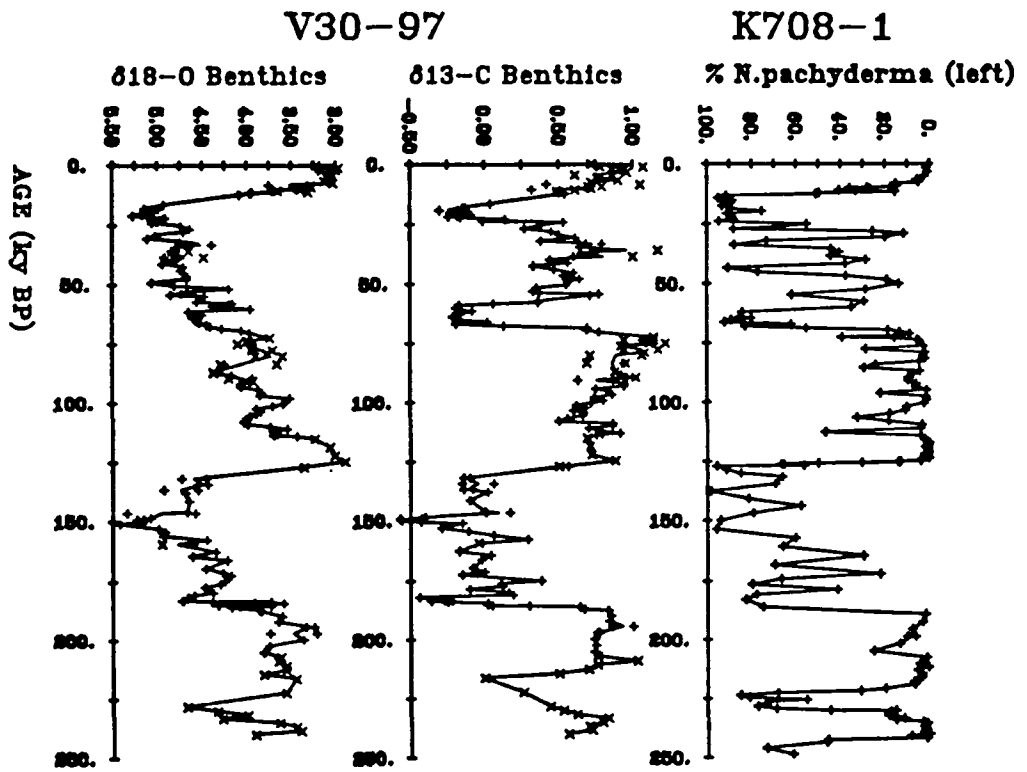

Fig. 1. Comparison of North Atlantic benthic isotope records from core V30-97 (41°00'N, 32°56'W, 3371m) to polar front migration recorded by % N. pachyderma (left-coiling) in core K708-1 (50°00'N, 23°45'W, 4053m). LEFT: For $\delta^{18}O$, which records mostly ice volume, +=U. peregrina and X=C. wuellerstorfii + 0.64. CENTER: For $\delta^{13}C$, which records both changes in global carbon budget and water-mass nutrients, +=U. peregrina + 0.90 and X=C. wuellerstorfii. RIGHT: High % N. pachyderma (left-coiling) indicates a southern position for the North Atlantic polar front in glacial time. Strong correlation between polar front position (right) and benthic $\delta^{13}C$ (center) suggests a direct link between surface- and deep-water oceanographic changes.

Three possible scenarios may explain the glacial pattern. First, if NADW formation stopped entirely, the deep Atlantic would be flooded with a single southern-source water mass. Our $\delta^{18}O$ data argue against this (Mix and Fairbanks, in press). Tropical Atlantic cores record deep-water temperatures higher (and/or salinities lower) than in the high-latitude North Atlantic. At least two deep-water masses were present in the glacial Atlantic.

Second, NADW flux may have been reduced, but not eliminated, with no charge in preformed nutrients of NADW. Third, NADW flux may have remained constant, but its preformed nutrients may have increased in glacial time as a response to heavy sea-ice cover and southern position of the polar front. A modern analogue for this scenario is the Weddell Sea.

A possible constraint is $\delta^{13}C$ in planktonic foraminifera from north of the polar front. This signal covaries with the deep ocean signal, which may suggest that preformed nutrients in NADW are not constant (Mix and Fairbanks, in press). Choosing between the second and third hypothese, however, depends on the mechanism for the global deep-ocean signal. If the ^{13}C budget reflects the transfer of forest biomass to the ocean in glacial time (Shackleton, 1977), polar planktonics reflects this global signal, and the second scenario is correct. If ^{13}C is linked to PO_4 in shelf organics (Broecker, 1981), polar planktonics reflect preformed nutrients in NADW, and the third scenario is correct. The truth probably lies somewhere in the middle, and both effects operate.

CHANGES IN THE HIGH-LATITUDE OCEAN AS POSSIBLE CAUSES OF ATMOSPHERIC CO_2 VARIATIONS

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Measurements on air enclosed in old polar ice have indicated that the atmospheric CO_2 concentration was ca. 50-70 ppm lower in late glacial times than during the Holocene (e.g., Neftel et al., 1982).

Similar measurements have been performed on samples from a Greenland ice core, dating ca. 30,000-40,000 B.P., and have yielded evidence of several CO_2 oscillations with an amplitude of ca. 50 ppm. Each change lasted on the order of a few centuries (Stauffer et al., 1984). A mechanism by which circulation changes in the high-latitude ocean could lead to rapid variations in atmospheric CO_2 is proposed (Siegenthaler and Wenk, 1984; Wenk and Siegenthaler, 1985). The atmospheric CO_2 level is controlled mainly by the physico-chemical properties of the surface ocean. Modification of the chemistry of the whole ocean seems to be too slow to explain these rapid variations. Marine biology exerts a strong influence on the $\sum CO_2$ and alkalinity of surface water, thus influencing the biological productivity. This is not the case in the Antarctic Ocean where nutrients are abundant. A slowing down of the vertical mixing would imply a smaller upward flux of $\sum CO_2$ and nutrients. Assuming constant productivity, $\sum CO_2$ and nutrients would be more completely used which would imply lower pCO_2 in these high-latitude surface waters. In areas with a warm surface, a slowing down of the circulation would not have a direct impact on pCO_2 because productivity would automatically decrease by the same factor as the upwelling rate of nutrients. Studies with a simple box model of the ocean-atmosphere system suggest that a sudden decrease by a factor of 2 of the water exchange between the surface and deep sea in high latitudes could lead to a CO_2 decrease of ca. 40-50 ppm with a time constant of ca. 200 years. Deep-sea sediment studies indicate rapid changes in the high-latitude surface conditions of the North Atlantic and the Antarctic Oceans at the end of the last glaciation. Studies of carbon isotope ratios should help us to ascertain whether this proposed mechanism was indeed responsible for the CO_2 variation.

The model calculations indicate that the high-latitude ocean is generally important for the atmospheric CO_2 since it is in rapid exchange with the large reservoir of the deep sea. CO_2 is transferred via the atmosphere between the different oceanic regions, so that pCO_2 in low latitudes tends to be similar to that of high latitudes.

Topics of Research

CO_2; climate:

- Vertical circulation and mixing between surface and deeper waters: mean rate of exchange; frequency of deep water formation events;

- Modification of surface properties in high latitudes by interaction with the atmosphere (heat and gas exchange) and with the marine biology (carbonate chemistry, nutrients);

- Possible applications of remote sensing: wind stress (important for gas exchange); biological productivity.

Ocean modeling:

- Effect of changed surface conditions (temperature, sea ice, wind stress) during Glacial on deep water formation;

- Effect of changed surface condition on thermocline processes, e.g. relation between stability and apparent vertical eddy diffusivity;

- Effect of changes in the North Atlantic on the other oceans, especially the Antarctic Ocean.

Paleoclimatic studies:

- Antarctic sediment studies: changes of surface conditions and of deep water formation; synchroneity with North Atlantic changes;

- Rapid variations during glacial periods: continental evidence from Europe and other continents, comparison with Greenland ice cores and deep-sea sediments.

OCEAN MODELING OF THE NORTH ATLANTIC

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A number of considerations make the North Atlantic Ocean a preferred part of the World Ocean to study. These are indicated in Table 1. The North Atlantic is especially attractive from a modeling viewpoint because a relatively complete set of processes are present in an ocean basin which is tractable for modelling and well-suited for satellite data acquisition. The understanding derived from studying the North Atlantic can be tested against present and past climates, generalized to the World Ocean, and applied to the prediction of future climates.

TABLE 1. Why Study the North Atlantic Ocean?

I. Practical Considerations
   - A. Focal point of climatic change
   - B. Site of nuclear waste disposal
   - C. Area of heavy shipping
   - D. Strategic importance
   - E. Biological productivity

II. Logistical Considerations
   - A. Much already known about the circulation
   - B. Many investigations focused there
   - C. Easier to observe than distant oceans
   - D. Amenable to remote sensing
   - E. Suitable for satellite data relay
   - F. Small enough to model

III. Physical Considerations
   - A. Region of diverse oceanic processes
     1. Mixed layer physics
     2. Sea-ice dynamics and thermodynamics
     3. Bottom water formation in high-latitude seas
     4. Marginal input of highly saline water
     5. Seasonal production of intermediate water
     6. Mixing along isopycnals
     7. Constraints on potential vorticity
     8. Unstable western boundary current
     9. Unstable mid-ocean flows
     10. Western boundary undercurrent
     11. Heat transport cross the equator
     12. Transient response to tracers
   - B. Understanding generalizes to the global ocean

Present modeling of the North Atlantic is inadequate and can be improved in a number of ways. Table 2 lists a number of important physical processes in five categories from the viewpoints of how they are treated in isolation, how they are usually represented in present ocean basin models, and how they may be better represented in future models. In the first two categories of vertical boundary processes and internal vertical mixing, parameterizations exist which can easily be incorporated into models and which will have important effects on the simulated structure of the North Atlantic. For the third category (mesoscale eddy effects), adequate parameterizations do not exist; but the order of magnitude of the effects is known from observational and process-model studies. A horizontal grid spacing of 100 km or less is required to allow parameterizations with this order of magnitude, as well as to resolve the time-averaged ocean fields (Fig. 1). Existing simulations with this gridsize and constant eddy diffusion coefficients show some success in reproducing oceanic phenomena that have previously been misrepresented in coarse-grid models. However, improved parameterizations are needed, since eddy resolving studies of the North Atlantic are computationally unfeasible, except for a few key simulations. In the fourth category of Table 2, improvements are suggested by way of increased vertical resolution and by the requirement that lateral mixing due to eddies takes place on isopycnal surfaces. Model incorporation of the latter phenomena is underway. In the fifth category of miscellaneous high-latitude processes, formulations for the treatment of sea ice are available for use. However, the treatment of gravitational instability, which is crucial to deepwater formation in the Atlantic Ocean, will require additional refinements to account for the unresolved physics of chimney formations in the open ocean and buoyant plumes near ocean boundaries.

It is suggested that a concentrated effort to understand the North Atlantic, involving the use and close interaction of conventional oceanographic measurements, satellite observations, and improved numerical models, could significantly advance our understanding of the oceanic component of the climate system.

TABLE 2. Ocean physical phenomena and methods of modeling them.

	Physical Phenomenon	Process Model Approach	Usual Climatic Model Approach	Improved Climatic Approach
upper & lower boundary processes	surface oceanic mixing	higher-order closure	constant-thickness mixed layer	prognostic-thickness bulk mixed layer (Heald & Kim, 1979)
	surface ageostrophic transport	vertically resolved	Ekman transport in upper level	Ekman transport in bulk mixed layer
	bottom boundary layer	vertically resolved	free-slip	quadratic drag law (Weatherly, 1972)
internal vertical mixing	vertical shearing instability	vertically resolved		
	double-diffusive effects	vertically resolved	$K_{H,M}$ = const.	$K_{H,M} = f_n(Ri,N)$
	small-scale vertical mixing	vertically resolved		(Pacanowski & Philander, 1981; Sarmiento et al., 1976)
effects of mesoscale eddies	barotropic instability		$A_M = 10^6 m^2 s^{-1}$ $A_H = 2x10^3 m^2 s^{-1}$	$A_M = A_H = 0(10^3 m^2 s^{-1})$
	baroclinic instabiltiy	horizontally resolved with Δx = 20 km	" "	$A_M = A_H = 0(10^3 m^2 s^{-1})$
	vertical form drag by eddies		K_M = const.	$K_M = f_n(\nabla T, \nabla u)$
large-scale transports	large-scale advection	done by time-mean currents	poorly resolved by 500 km, 5 level grid	moderately resolved by 100 km, 10 level grid
	mixing along isopycnals	isentropic coordinates	z coordinates	diffusion along isopycnals (Redi, 1982)
	interaction with topography	sigma coordinates	flat bottom	variable number of levels (Bryan, 1969)
high-latitude processes	density relation	Knudsen formula	linear/quadratic formula	higher order polynomial (Bryan & Cox, 1972)
	gravitational instabiltiy	plume models	convective adjustment	modified convective adjustment
	sea-ice dynamics	viscous/elastic/ plastic	no motion	empirical ice motion (Thorndike & Colony, 1982)
	ice thermodynamics	storage of sensible and latent heat	no heat storage	storage of sensible and latent heat (Semtner, 1976)

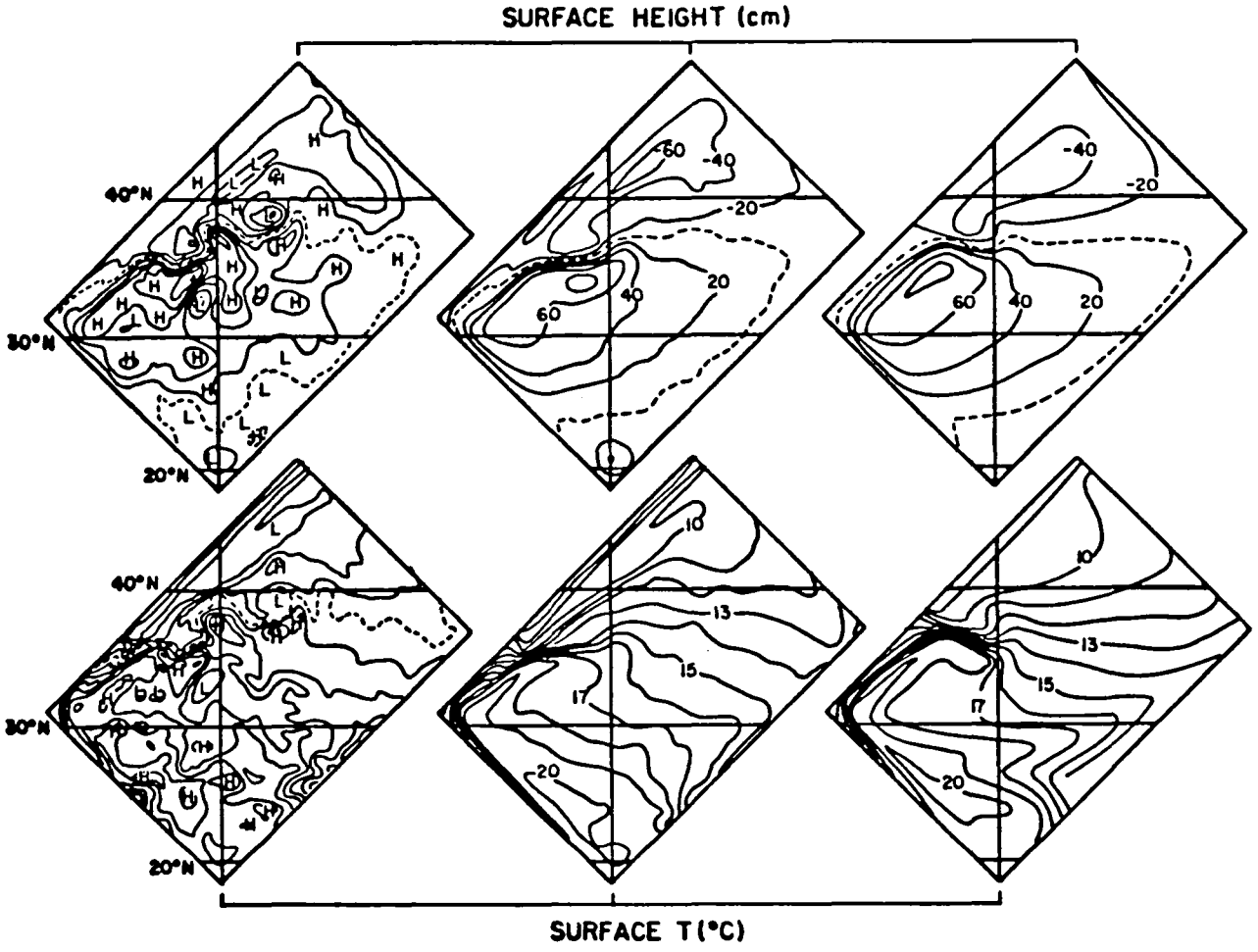

Fig. 1. Surface height and temperature fields from the study of Semtner and Mintz (1977). On the left are instantaneous fields from an eddy resolving simulation with 37 km gridsize. The time-averaged fields from this experiment are shown in the center. On the right are the steady fields from a coarser grid experiment (75 km gridsize) with constant eddy diffusion coefficients ($A_M = A_H = 10^3$ m^2 s^{-1}).

NUMERICAL SIMULATION OF NORTH ATLANTIC SEA ICE VARIABILITY, 1951-1980

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A two-level dynamic-thermodynamic sea ice model (Hibler, 1979) is used to simulate the growth, drift and decay of sea ice in the Northern Hemisphere during a 30-year period, 1951-1980. The model is run with a daily timestep on a 222 km grid (Fig. 1) and is forced by interanually varying fields of geostrophic wind and temperature-derived thermodynamic fluxes. The objective is a quantitative description of large-scale sea ice variability in terms of the dynamic and thermodynamic processes responsible for the fluctuations, especially in the North Atlantic where sea ice represents a substantial input of fresh water.

The fields of ice velocity and thickness contain strong seasonal as well as interannual variability. The mean drift pattern results in thicknesses of 4-5 m offshore of northern Canada and Greenland, while winter thicknesses of ~2 m are typical of Alaskan, Eurasian and East Greenland coastal waters. The 30-year mean fields are characterized by excessive ice in the North Atlantic during winter and by a summer retreat that is more rapid than observed. The excess of winter ice in the North Atlantic is due primarily to the omission of horizontal heat transport and deep convection in the ocean. Annual net growth ranges from 0.1 to 0.6 m over much of the Arctic Basin and Baffin Bay, while annual net melt of 0.5-1.5 m occurs in the North Atlantic marginal ice zone (Fig. 2).

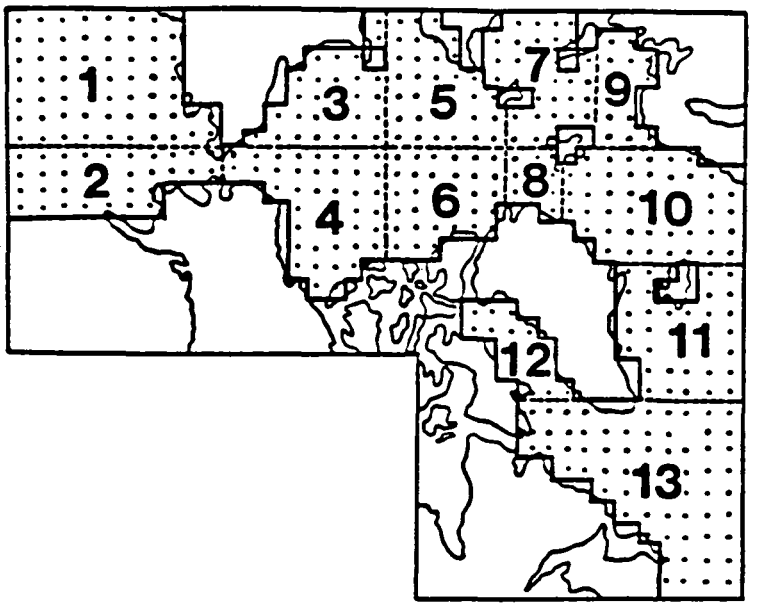

Fig. 1. The model domain and the 13 regions for of net which the mass budget statistics were evaluated.

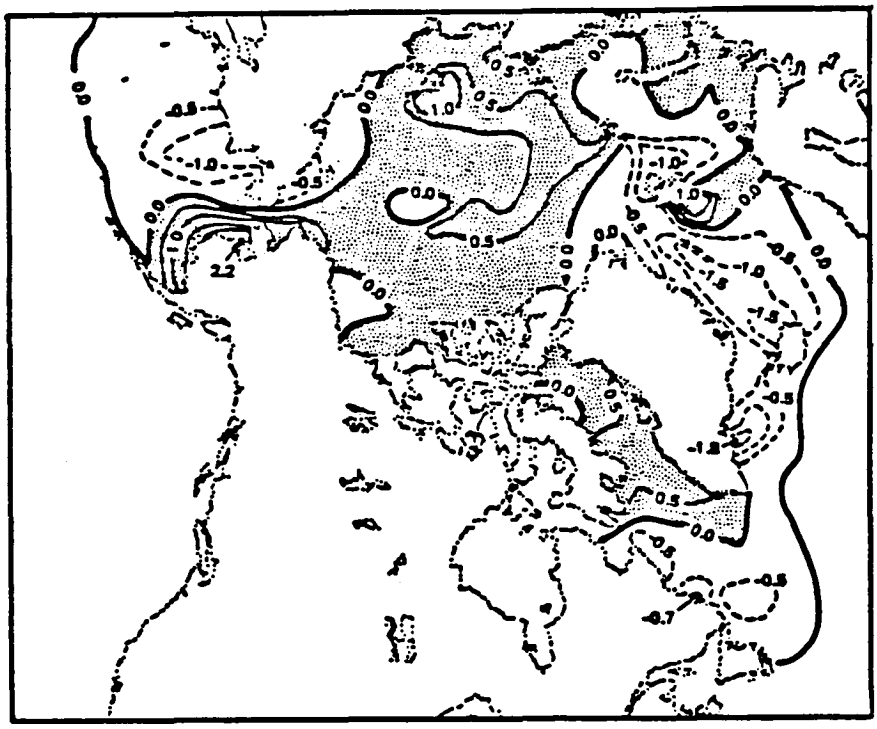

Fig. 2. 30-year mean distribution annual ice growth (m).

Despite the biases in the mean fields, the simulated interannual fluctuations correlate at 0.4-0.9 with the corresponding observed fluctuations in individual sectors. In a comparison of results based on different data sources, closer agreement with observed fluctuations was obtained with the thermodynamic fields derived from the NASA GISS temperature grids than from the grids compiled in the USSR. The simulated velocities show no bias but considerable scatter relative to the drift of the Arctic bouys during 1979 and 1980 in the central Arctic and the Greenland Sea.

An analysis of the regional mass budget indicates that the normal seasonal cycle is controlled primarily by thermodynamic processes, but that thickness anomalies in much of the Arctic are attributable primarily to dynamic processes during winter, spring and autumn (Fig. 3). Thermodynamic processes contribute more strongly to summer anomalies and to anomalies near the ice edge.

The tendency for thickness anomalies to be advected by the pattern of mean drift is apparent in multiseason lag correlations involving subregions of the Arctic Basin and the peripheral seas. Anomalies east of Greenland (Region 8 in Figs. 1 and 3) are highly correlated with anomalies of ice transport into the Fram Strait. The simulated outflows of ice mass through the Fram and Bering Straits vary by factors of 2-3 in successive years (Fig. 4). The mean ocean currents prescribed in the simulations account for 20-30% of the simulated outflow in the Fram Strait.

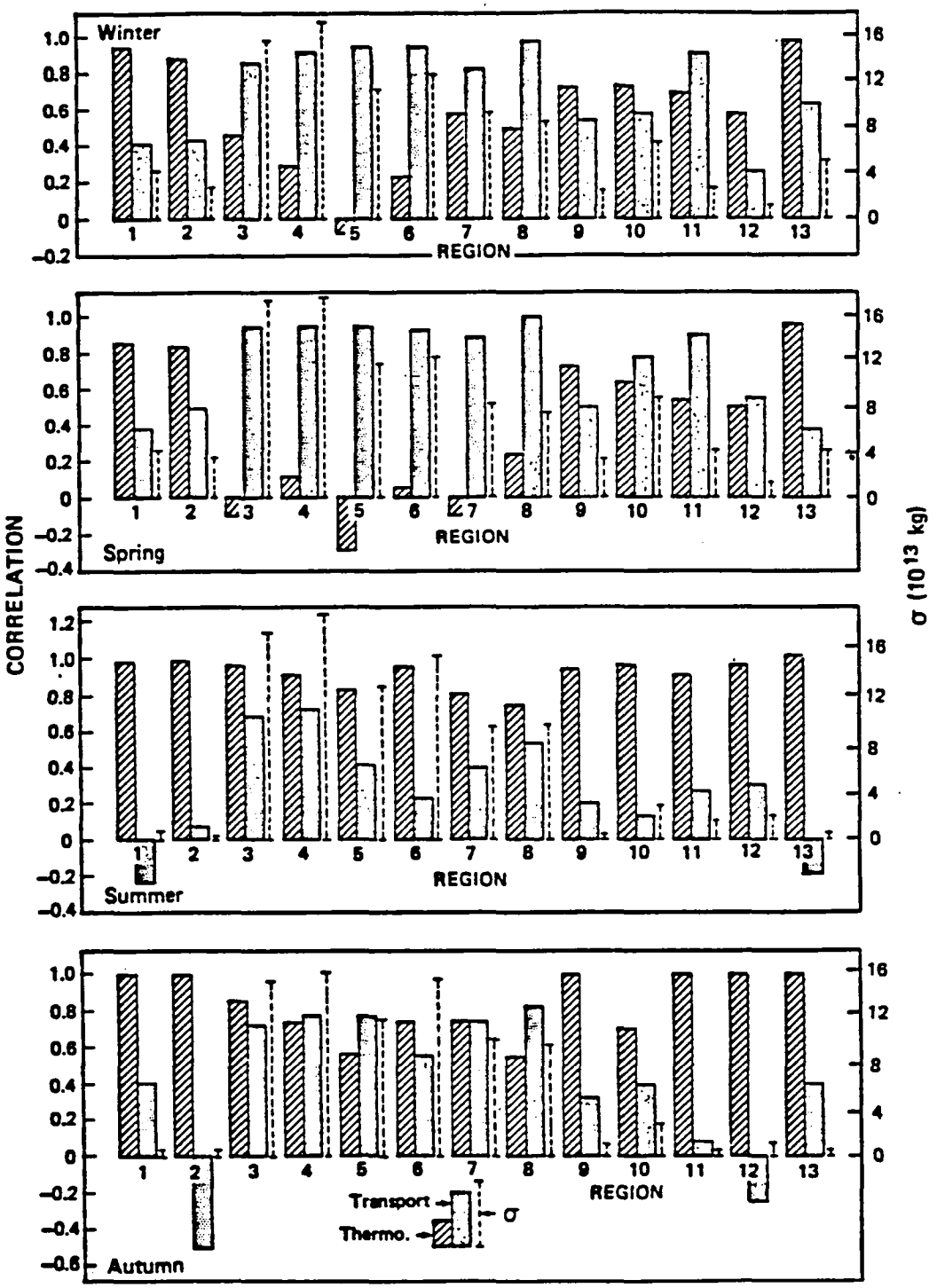

Fig. 3. Correlations between 2-month changes of regional anomalies of mass and departures from normal growth/melt (hatched bars) or advective flux convergence (solid bars) in each region. Results are shown for all 2-month periods ending in winter (December-February), spring (March-May), summer (June-August), and autumn (September-November). Standard deviation (10^{13} kg) of monthly mass anomalies in each region are shown adjacent to correlation bars.

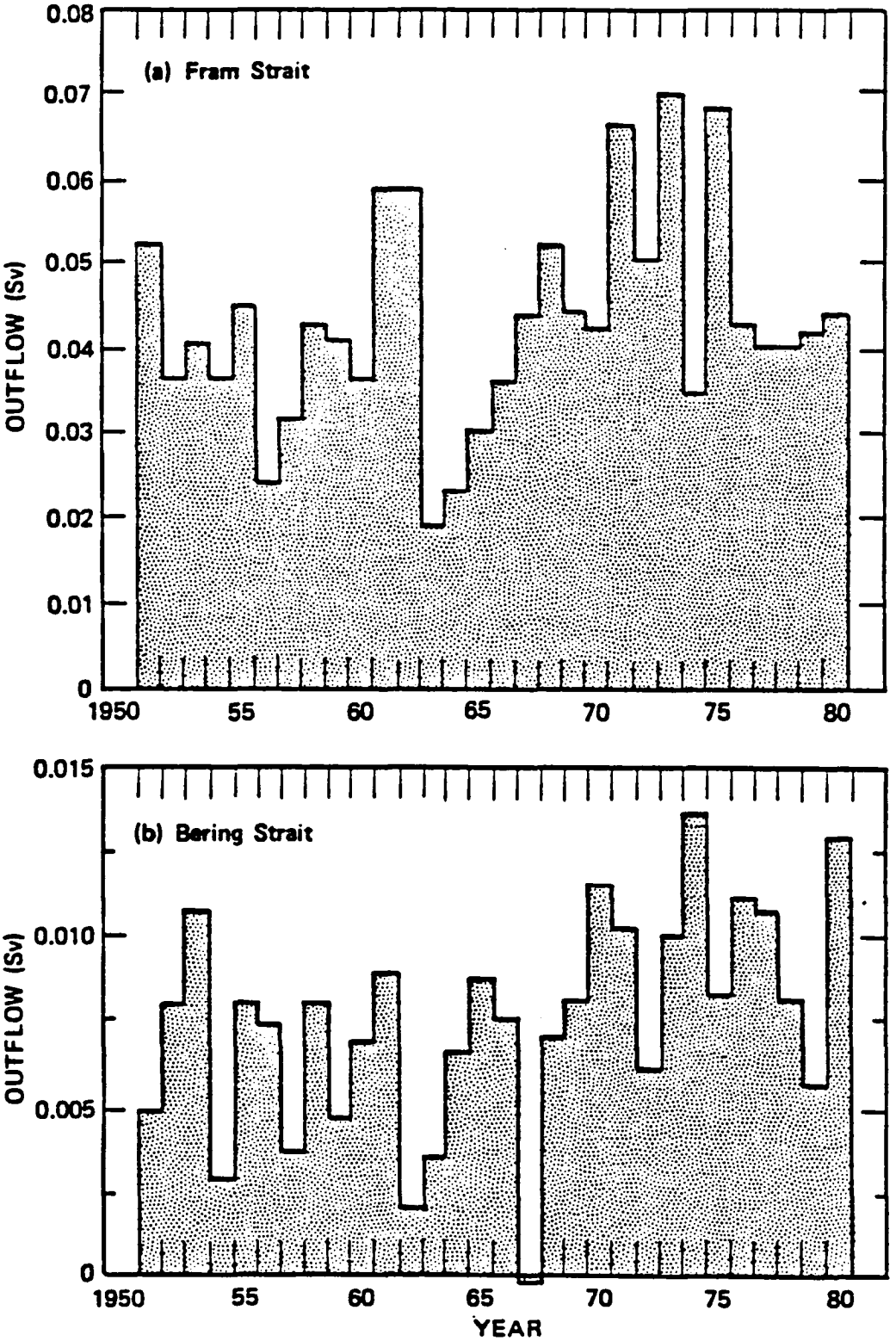

Fig. 4. Annual ice outflow (S_V) through (a) Fram Strait and (b) Bering Strait. Positive values represent new outflow, negative value represents net inflow during January-December, inclusive.

POSSIBLE SEA ICE IMPACTS ON OCEANIC DEEP CONVECTION

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Many regions of the world ocean known or suspected to have deep convection are sea-ice covered for at least a portion of the annual cycle (Fig. 1). As this suggests that sea ice might have some impact on generating or maintaining this phenomenon, several mechanisms by which sea ice could exert an influence are presented in the following paragraphs.

Sea ice formation could be a direct causal factor in deep convection by providing the surface density increase necessary to initiate the convective overturning (Fig. 2). As sea ice forms, either by ice accretion or by in situ ice formation in open water or in lead areas between ice floes, salt is rejected to the underlying water. This increases the water salinity, thereby increasing water density in the mixed layer under the ice. A sufficient increase in density will lead to mixing with deeper waters, and perhaps to deep convection or even bottom water formation. Observations are needed to establish whether this process is actually occurring; it is most likely in regions with extensive ice formation and a relatively unstable oceanic density structure.

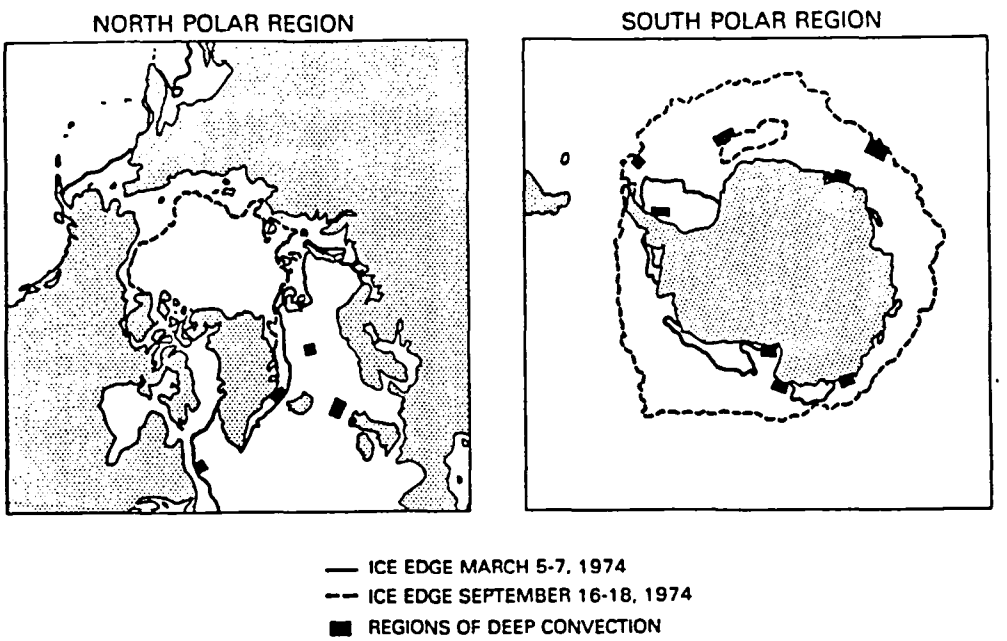

Fig. 1. The approximate seasonal range of the ice edge location in both the north and south polar regions, plus the approximate locations of oceanic deep convection. The ice edge is determined from the Electrically Scanning Microwave Radiometer on board NASA's Nimbus 5 satellite, using the 193 K brightness temperature contour, while the locations of deep convection are approximated from Killworth (1983; Figs. 2 and 10).

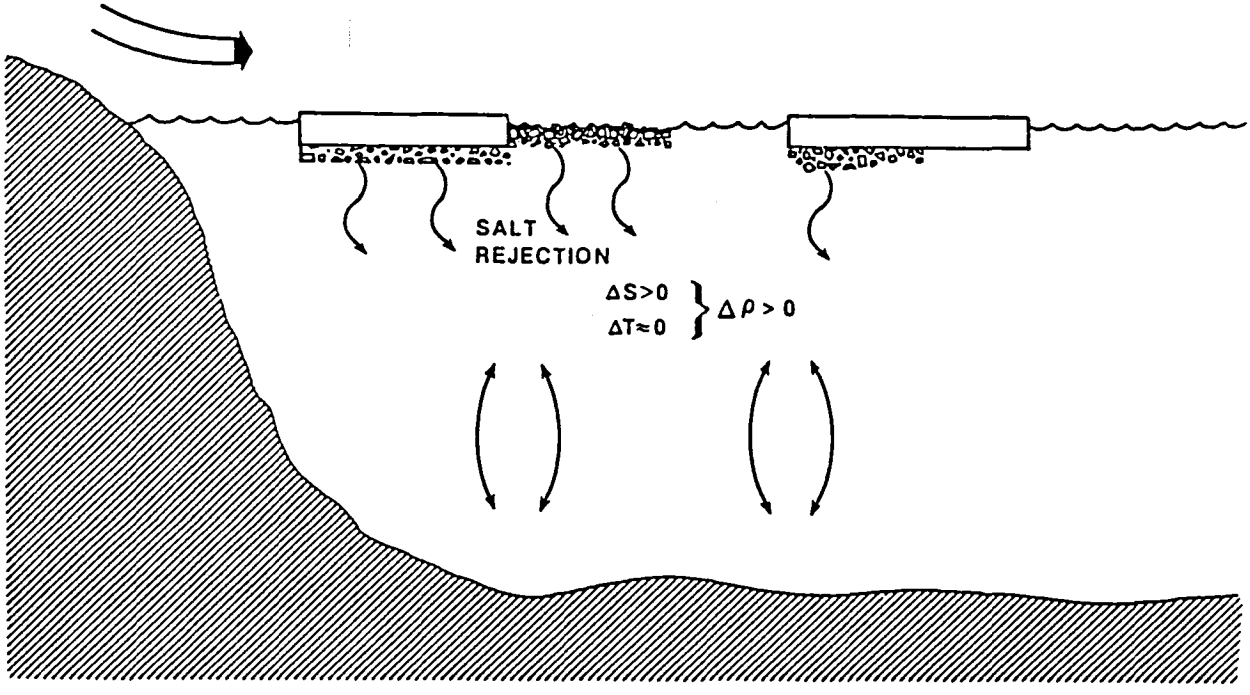

Fig. 2. Schematic representation of a possible mechanism for deep convection, with sea ice formation the primary causal factor. Salt rejection during ice formation induces convection by increasing the surface water density.

In regions and periods with deep convection just seaward of the ice edge, the presence of the ice, rather than its formation, could contribute by intensifying the ocean/atmosphere contrast. Specifically, air flowing off the Greenland or Antarctic ice caps, already cold and dry, is further cooled radiatively as it flows over the adjacent sea ice, so that the temperature contrast between ocean and atmosphere is even greater once the air reaches open water either in a polynya or beyond the ice edge (Fig. 3). The large resulting heat and evaporative fluxes from ocean to atmosphere decrease ocean temperatures and increase ocean salinities. Both of these adjustments increase surface water densities, and, with sufficient density increase, can lead to mixing with deep waters and perhaps to deep convection and even bottom water formation. In this case sea ice has an indirect role, contributing to the creation of the atmospheric conditions which allow the fundamental ocean/atmosphere mechanism to operate.

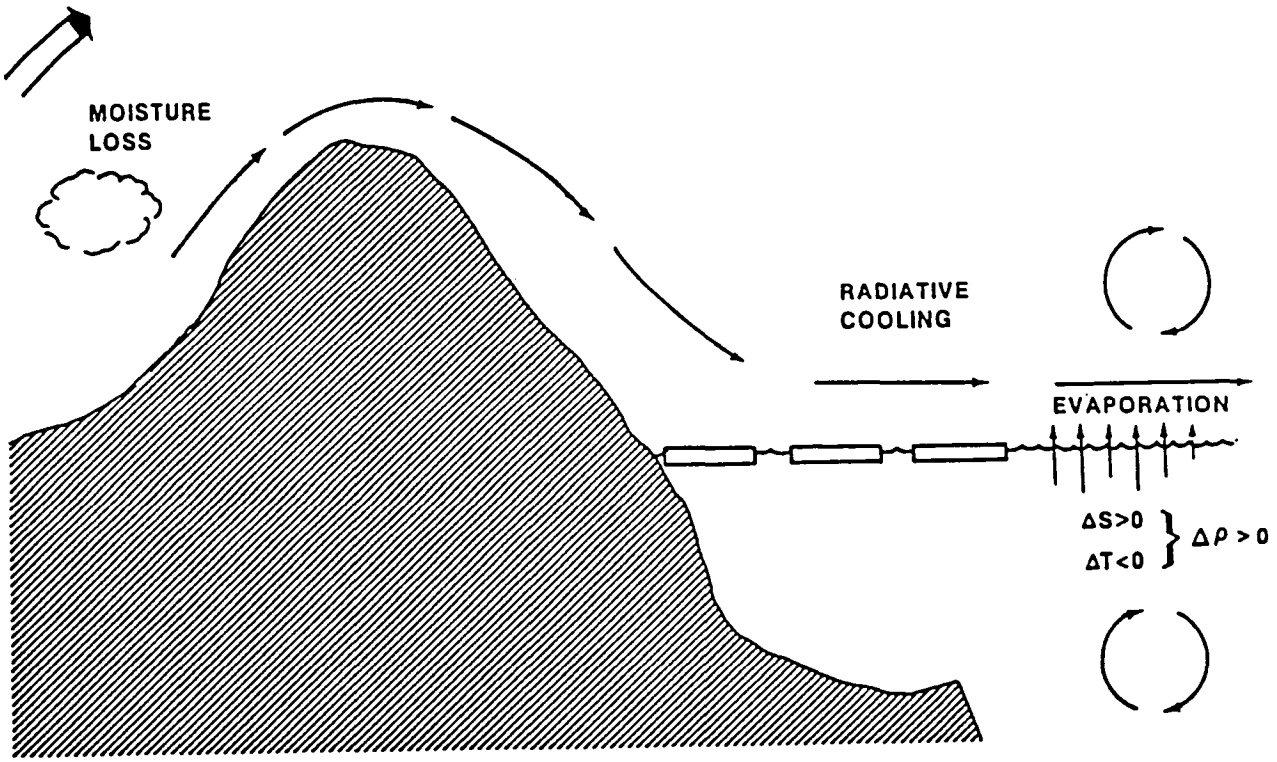

Fig. 3. Schematic representation of a possible mechanism for deep convection, with the presence of sea ice a contributing factor. Radiative cooling as air passes over sea ice intensifies the air/sea contrast beyond the ice edge.

If the water density increase from salt rejection during sea ice formation is insufficient to cause deep convection and the primary mechanism instead depends upon direct ocean/atmosphere contact, then the presence of sea ice has a strong negative role in addition to the positive role indicated in the previous paragraph for deep convection just beyond the ice edge. Within the pack, sea ice would prevent deep convection by the effective insulation it provides between ocean and atmosphere (Maykut (1978) indicates that heat input to the atmosphere is reduced by one or two orders of magnitude in the presence of an ice cover of one meter or more in thickness). Hence a retreat of the ice edge in a warmer climatic period could lead to deep convection in regions which are currently ice covered, and an advance of the ice edge in colder periods would eliminate deep convection in those regions currently just seaward of the ice edge.

To conclude, whether or not sea ice formation directly contributes to deep convection through the rejection of salt, the presence of sea ice almost certainly has an impact on the location of deep convection in the polar oceans and thereby presumably influences the changes in deep ocean circulation from one set of climatic conditions to another.

SEA ICE, CLIMATE AND FRAM STRAIT

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When sea ice is formed the albedo of the ocean surface increases from its open water value of about 0.1 to a value as high as 0.8. This albedo change affects the radiation balance and thus has the potential to alter climate. Sea ice also partially seals off the ocean from the atmosphere, reducing the exchange of gases such as carbon dioxide. This is another possible mechanism by which climate might be affected.

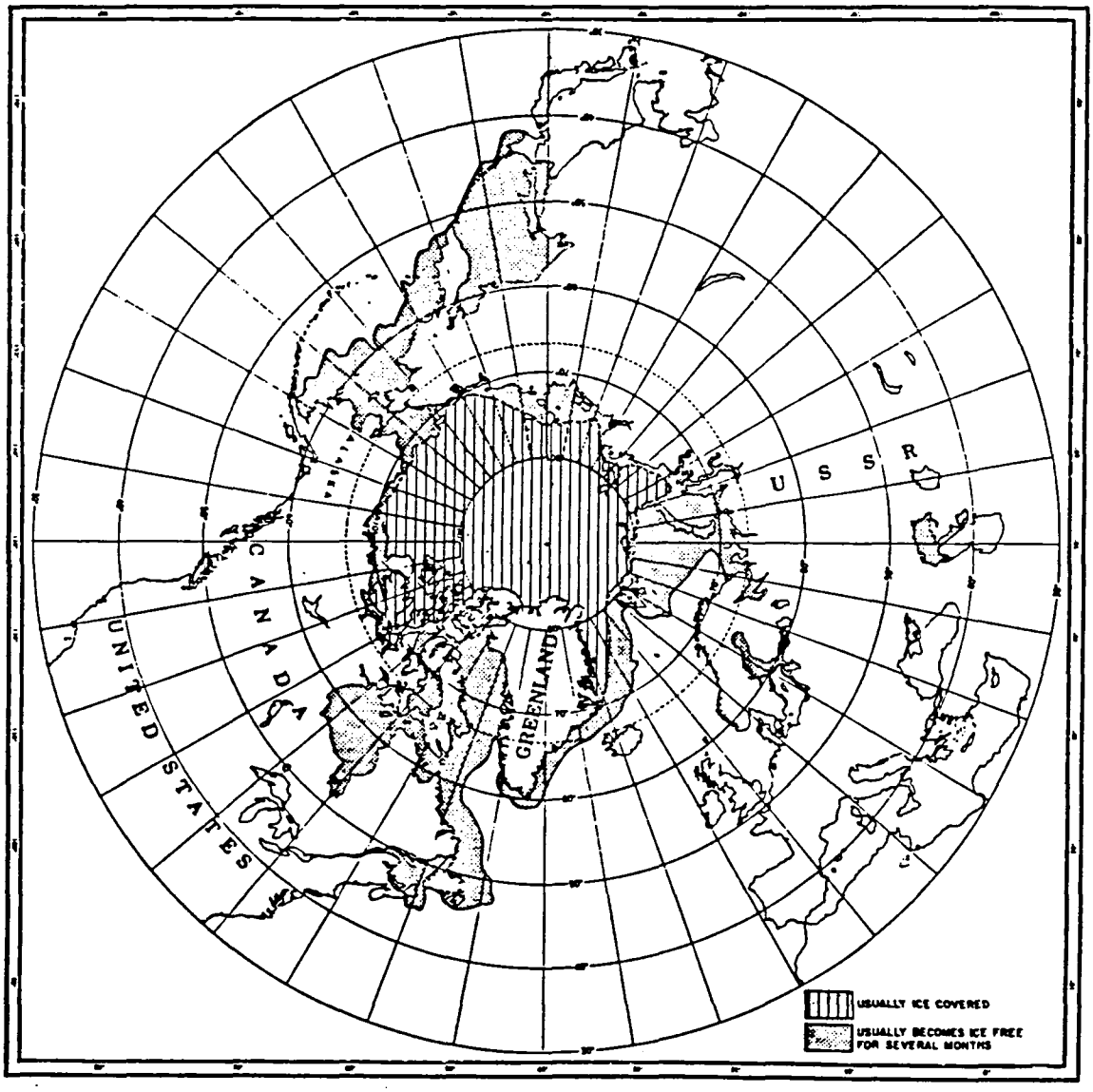

Fig. 1. Maximum and minimum extent of sea ice in the Northern Hemisphere.

The central Arctic Ocean is covered throughout the year with polar pack ice about 3 m in thickness. However, the shelf areas and the surrounding seas such as the Barents and Bering are covered only seasonally with ice about 1 m in thickness (Fig. 1). In the absence of land masses and ocean currents sea ice might be expected to have a zonally uniform distribution. But the actual distribution is highly irregular in southward extent. The most striking deviation occurs east of Greenland where a tongue of sea ice extends 2,000 km southward of Fram Strait, the passage between Greenland and Spitsbergen, in winter and half that distance in summer. The seas just east of this tongue from Fram Strait southward are free of ice throughout the year. Sea ice is exported from the Arctic Ocean into the Greenland Sea and North Atlantic (Fig. 2). This Strait and the surrounding region have been studied by two recent research programs. The FRAM Expeditions from 1979-82 investigated the oceanographic conditions north of the Strait with the aid of manned drifting stations and helicopters. The Marginal Ice Zone Experiment (MIZEX 83-84) is an international, multidisciplinary study of processes controlling the edge of the ice pack in that area including the interactions between sea, air and ice.

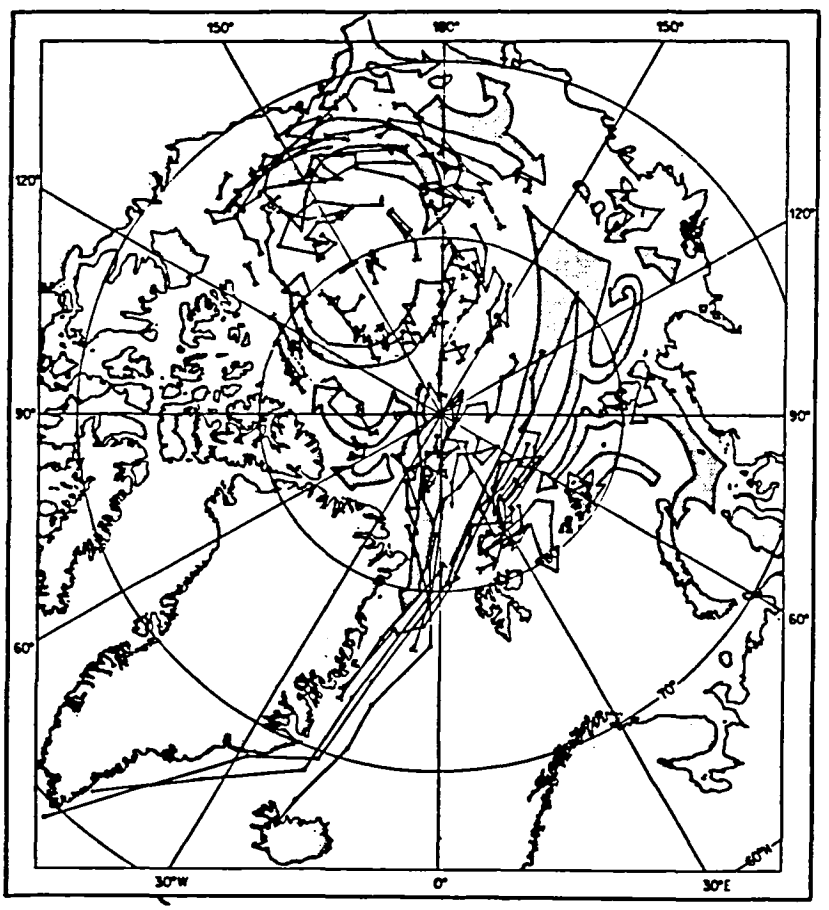

Fig. 2. Drift paths of 45 air-dropped data buoys between January 1979 and December 1981. The buoys are positioned by ARGOS several times a day to an accuracy better than 1 km. They contain quartz-oscillator pressure sensors whose readings are accurate to ±1 mb, with a drift of less than 0.2 mb per year (adapted from Untersteiner and Thorndike, 1982). The stippled arrows indicate an early (Gordienko, 1958) estimate of the mean circulation of the upper several hundred meters of water in the Arctic Basin. (From Air-Sea-Ice Research Program for the 1980's, APL-Univ. of Washington, 1983).

Sea ice can be driven by wind on its upper surface or ocean currents on its lower surface. In Fram Strait mean annual winds are southward and the ice moves southward, even on the Spitsbergen side where the currents are northward (Fig. 3). Oceanographic conditions are sharply contrasting across the Strait with the colder, less saline East Greenland Current (EGC) flowing southward on the western side and the warmer, more saline West Spitsbergen Current flowing northward on the east side. The combined effect of wind and the cold EGC carry ice far southward off the coast of Greenland. But off Spitsbergen the southward-moving ice is melted by waters with temperatures above freezing. Oceanographic surveys by helicopter during FRAM experiments showed a horizontal boundary layer beneath the ice in which the WSC is melting polar pack ice, reducing its thickness from 3 m to zero in a distance of 50 to 100 km. The export of sea ice from the Arctic Ocean to the North Atlantic will be affected by both wind and current regimes. Mean winds in this region may be considered a part of the polar anticyclone and would be altered if the global meridional temperature gradient changed. Ocean currents here are driven by the density contrast between the low-salinity Arctic Ocean fed by river runoff from northern Eurasia and the saline waters which have been carried northward from the tropical Atlantic. Any changes in discharge of the Siberian Rivers or transport of the Gulf Stream would also have an effect on ice discharge from the Arctic Ocean.

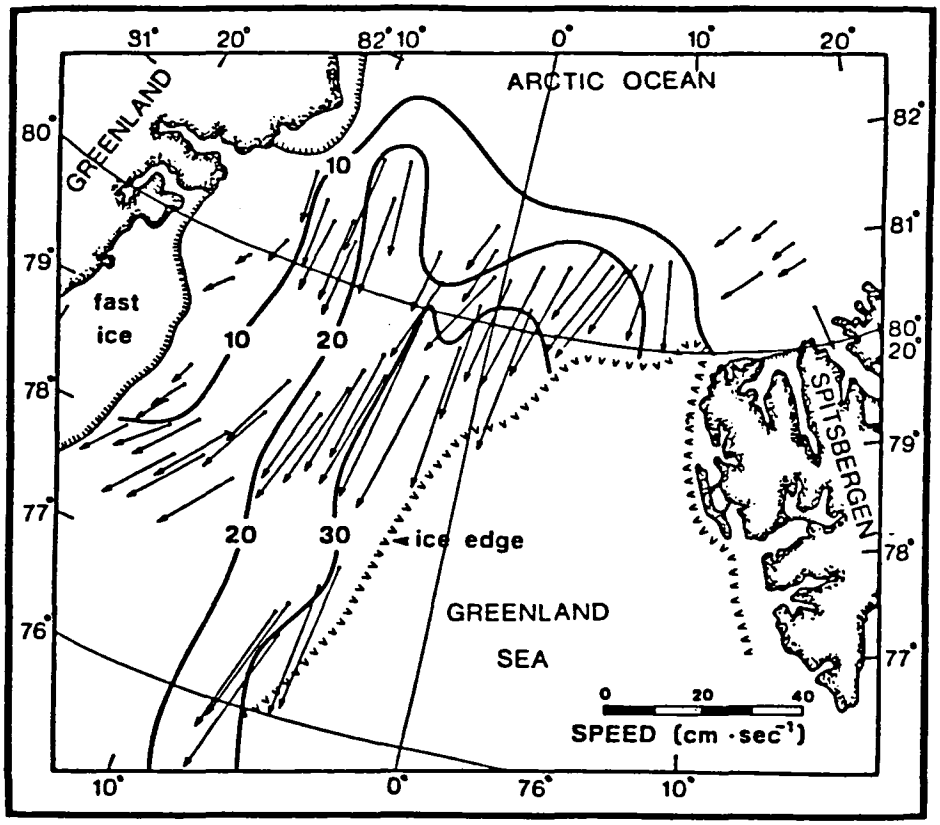

Fig. 3. Ice drift observations for 21 April - 7 May 1976. Heavy lines are isotachs cm sec^{-1}. (From Rey, 1982, p. 89)

GREENLAND ICE SHEET MASS BALANCE

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1. The mass balance equation

The mass balance equation, which expresses the law of mass conservation for a glacier, may be written:

$$\frac{\partial V}{\partial t} = Q_P - Q_M - Q_{CALF} \quad , \tag{1}$$

where V is the glacier volume, t is time, Q_P is the annual precipitation, Q_M and Q_{CALF} are the annual volume losses by melting and calving of icebergs, respectively. All volumes are expressed in terms of water equivalents. According to equation (1), the total water loss, $Q_L = Q_M + Q_{CALF}$ may be calculated as

$$Q_L = Q_P - \frac{\partial V}{\partial t} \quad , \tag{2}$$

i.e. as the total precipitation over the glacier minus the time rate of ice-volume increase.

2. Areal distributions and ice volumes

The following table shows the distribution of glacierized and ice-free land in Greenland (Holtzscherer and Bauer, 1954).

	Area (km^2)	%
Greenland ice sheet	1.730.000	79
Local glaciers	80.000	4
Ice-free land	380.000	17
Total	2.190.000	100

For the Greenland ice sheet, the accumulation and ablation areas constitute 1.440.000 km^2 and 290.000 km^2, i.e. 84% and 16% respectively, of the total ice sheet area. The volume of the Greenland ice sheet has been estimated at 2.400.000 km^3 of water equivalent (Holtzscherer and Bauer, 1954), whereas the volume of the local glaciers has been given by Weidick (1975b), as no more than 100.000 km^3.

3. Estimates of actual mass balance

Several estimates have been made for the actual gain, loss and total balance of the Greenland Ice Sheet. They have been summarized by Weidick (1984) as follows:

Accumulation	500 ± 100 km^3 of w. eqv.
Melting	295 ± 100 km^3 of w. eqv.
Calf-ice	260 ± 60 km^3 of w. eqv.

The estimates show, that the Greenland Ice Sheet is believed not to be greatly out of balance under the present climatic conditions. This is confirmed by observations of recent thickness changes of the ice sheet, which generally show average thinning-rates on the order of 0.2 m a^{-1} in parts of the ablation area of southern and central west Greenland (Bauer et al., 1968a; Weidick, 1968; Seckel, 1977) and thickening-rates on the order of 0.1 m a^{-1} in central Greenland (Seckel, 1977).

3.1. Actual precipitation-distribution

The map in Fig. 1 shows the actual distribution of precipitation over Greenland. The map is based on accumulation rate data for the ice sheet, obtained by firm stratigraphic methods (pit studies, ice corings) and precipitation records from coastal stations. Compared to previous compilations (e.g. Mock, 1967 and Radok et al., 1982) the data comprises results obtained by the Greenland Ice Sheet Program (Langway et al., 1984) which has supplied new information in particular as regards the southern Greenland Ice Sheet (Reeh and Clausen, in prep.). The main difference between Fig. 1 and previous maps (e.g. Mock, 1967) is, that the accumulation rate over the southwestern slope of the ice sheet between 65N and 69N shown in Fig. 1, is only about 50% of the corresponding values given by Mock (1967).

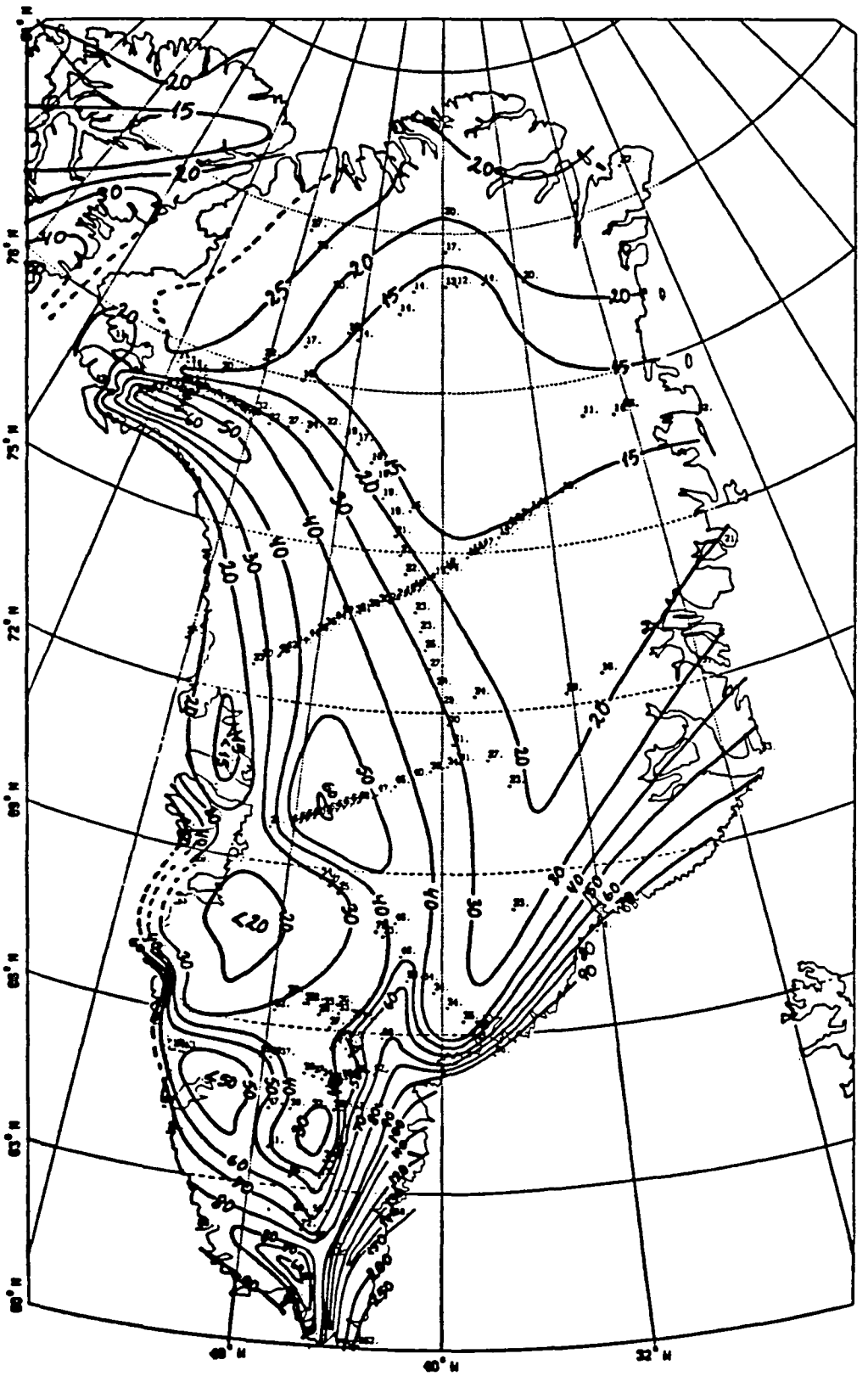

Fig. 1. Distribution of precipitation in Greenland (g cm^{-2} a^{-1}). Compiled by Reeh, 1984.

3.2. Melting from the Greenland Ice Sheet

According to Holtzsherer and Bauer (1954), the mean net balance (net ablation) over the ablation area is -1.1 m of w. eqv/year. (The net balance is defined as the net loss of ice-mass or -volume, i.e. as the accumulation minus the total melt.) This estimate is based on scattered observations from the ablation zone in west Greenland. In east- and north-Greenland very few ablation rates have been measured on the ice sheet. Net ablation rates are probably lower than has been previously estimated, in southeastern Greenland due to neglect of the large precipitation over the ice sheet margin (see Fig. 1) and in east central Greenland due to high elevations of the ice sheet margin.

3.3. Loss by calving of icebergs

The most extensive studies of iceberg discharge from the Greenland ice sheet are those of Bauer et al. (1968b) and Carbonnell and Bauer (1968), who determined the calf-ice production of the outflow glaciers calving into Disko Bugt and Umanak Fjord, west Greenland by means of repeated areal photography. The results of these two studies deviate by about 10% (90 km^3 of ice pr. year in 1957, and 102 km^3 of ice pr. year in 1964).

Based on the investigations mentioned above and scattered information about calving glaciers in southwest Greenland, Weidick and Olesen (1978) estimate the total calf-ice production from west Greenland glaciers between 60N and 71N to be 97 km^3 of water pr. year. As regards north and east Greenland very few observations of calving rates have been made. Olesen and Reeh (1969) give the calf-ice production from the northern part of the Scoresby Sund region to be 11 km^3 of ice per. year.

3.4. Present hydrological budget for Greenland

Assuming the Greenland Ice Sheet to be presently in balance with the accumulation rate distribution shown in Fig. 1 and assuming net ablation rates as the ones given in the Table below, the values for the various terms in the hydrological budget for Greenland can be estimated. The results are presented in the Table below, where the sectors refer to the division of the Greenland Ice Sheet shown in Fig. 2. The total calf-ice discharge for the various sectors can be distributed between the main outlet glaciers according to their catchment areas. The result of such a division is shown in Fig. 2. (Where calving rates have been observed, observed values are shown.)

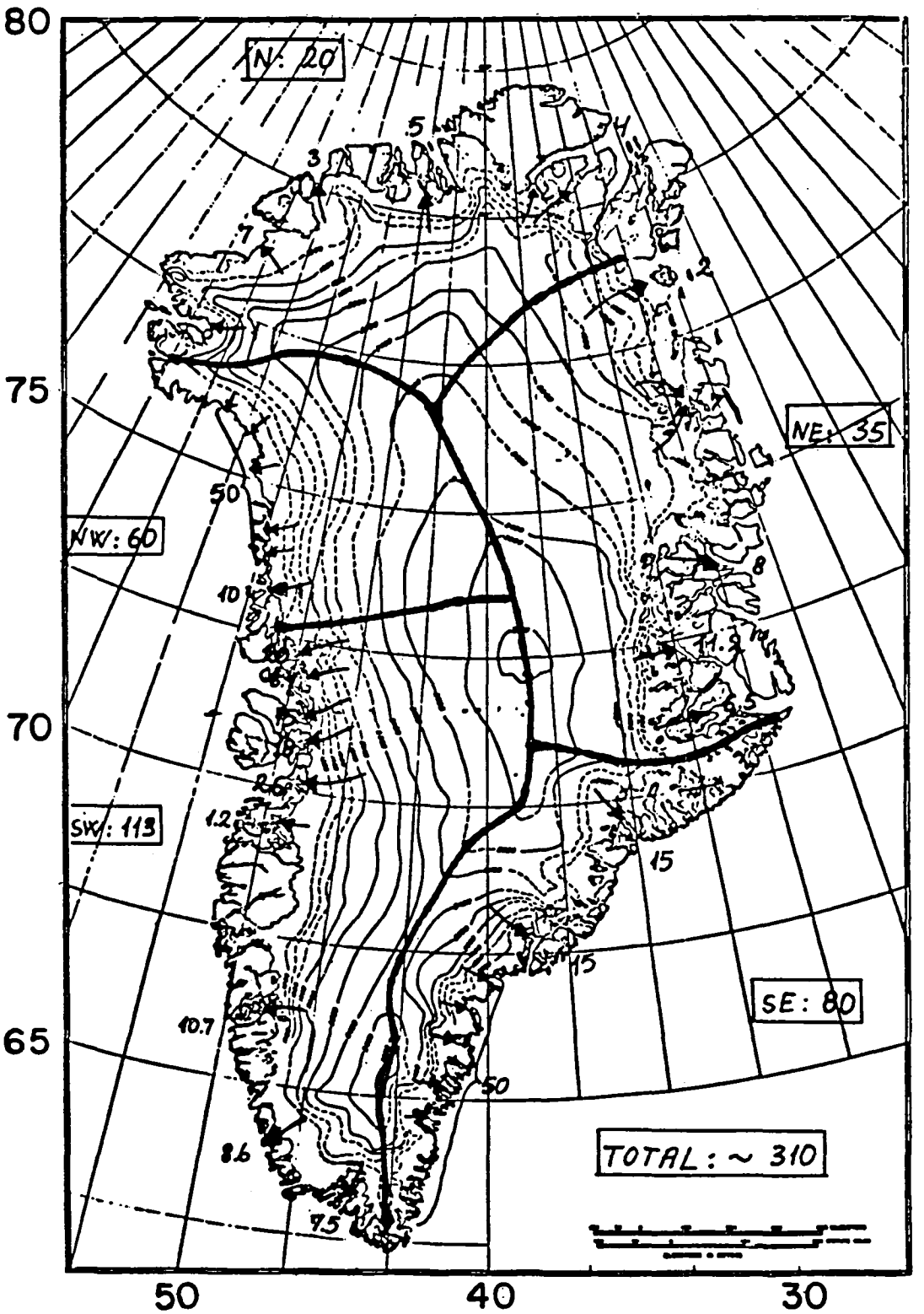

Fig. 2. Estimated calf-ice discharge from Greenland ice sheet in km^3 of water equv. per year. Compiled by Reeh, 1984.

Sector	$Q_{P,ACC}$ (km^3/a)	$Q_{P,AB}$ (km^3/a)	$Q_{P,IF}$ (km^3/a)	$Q_{P,TOT}$ (km^3/a)	$Q_{N,AB}$ (km^3/a)	$Q_{M,AB}$ (km^3/a)	Q_{CALF} (km^3/a)	$Q_{R,IF}$ (km^3/a)	Net-Ablation (km^3/a)
SW	178	27	34	239	60	87	118	34	1.0
NW	83	8	2	93	22	30	61	2	1.0
N	47	9	32	88	26	35	21	32	0.7
NE	74	11	29	114	38	49	36	29	0.7
SE	105	15	73	193	23	38	82	73	1.0
TOTAL	487	70	170	727	169	239	318	170	

$Q_{P,ACC}$ = precipitation over accumulation area of ice sheet
$Q_{P,AB}$ = precipitation over ablation area of ice sheet
$Q_{P,IF}$ = precipitation over ice-free land and local glaciers
$Q_{P,TOT} = Q_{P,ACC} + Q_{P,AB} + Q_{P,IF}$
$Q_{N,AB}$ = net balance (net ablation) over ablation zone
$Q_{M,AB} = Q_{P,AB} + Q_{N,AB}$ = total melt in ablation zone
Q_{CALF} = calf-ice discharge
$Q_{R,IF}$ = run-off from ice-free land and local glaciers

4. Temporal variations of Greenland mass balance

Analyses of ice cores from the Greenland Ice Sheet indicate (i) fairly constant accumulation rates back to 1400 B.P. (Reeh et al., 1978), (ii) up to 25% increased accumulation rates in the period 3000-7000 B.P. (Reeh et al., 1984, figures 7 and 8), and (iii) a more than 50% decrease in the accumulation rate in late Wisconsin time (Hammer et al., 1978).

Past ablation rates and calving rates cannot be deduced by direct measurements. However, volume changes of the ice sheet can be estimated from the history of ice margin fluctuations (Weidick, 1975b), see Fig. 3a, combined with ice sheet profile theory. Fig. 3b (from Weidick, 1975b), shows derived ice sheet volume changes since 14,000 B.P. Assuming the above mentioned temporal accumulation rate changes to be representative for the entire ice sheet, and deriving the rate of ice sheet volume change from Fig. 3b, the time history of the total water loss from the Greenland Ice Sheet can be calculated by means of equation (2). The results are presented in the following table.

Time (years B.P.)	Ice Sheet volume loss (km^3/a)	Accumulation on ice sheet (km^3/a)	Water loss from ice sheet (km^3/a)	Run-off from ice-free area (km^3/a)	Total water loss from Greenland (km^3/a)
12000	0	400	400	0	400
10000	150	430	580	20	600
8500	320	490	810	80	890
7000	100	690	790	190	980
5500	0	560	560	170	730
0	0	560	560	170	730

The table shows that since 12000 B.P. the annual loss of water from Greenland has ranged from 45% less to 33% more than the present water loss. These variations are surprisingly small, since the ice sheet in the same period has experienced a change in climate from glacial to interglacial conditions.

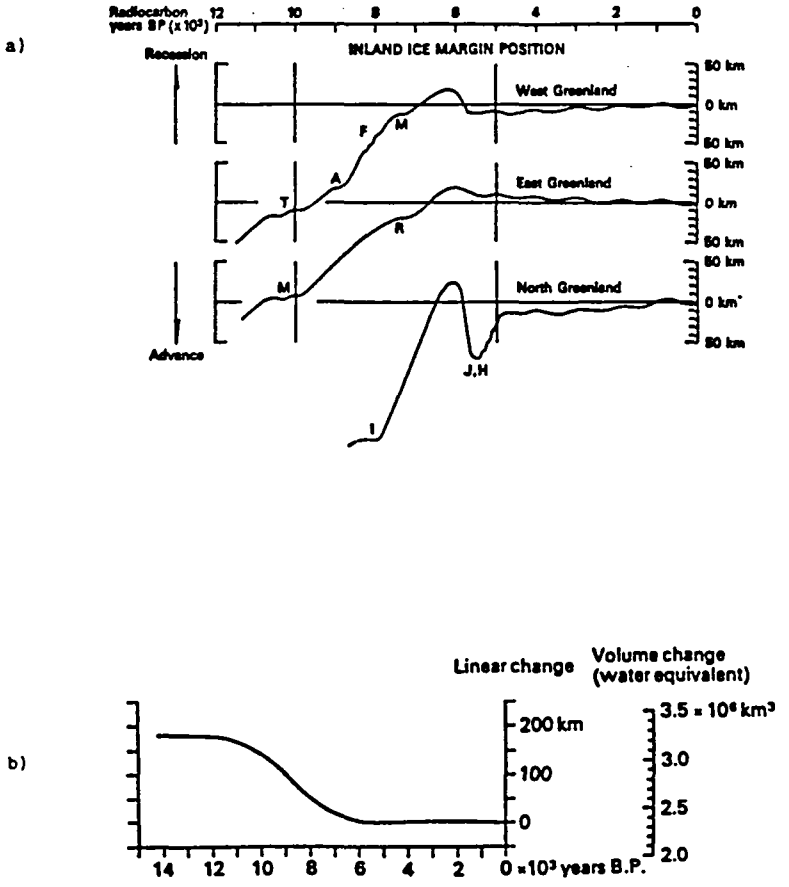

Fig. 3. a) Ice margin fluctuations in Greenland since 12000 B.P. (from Weidick, 1975b). b) Average recession of the Greenland Ice Sheet margin. The recession is also converted to volume shrinkage (from Weidick, 1975a).

POTENTIAL CLIMATIC EFFECTS ON THE GREENLAND ICE SHEET

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The Greenland Ice Sheet covers an area of 1.72×10^6 km^2 and contains approximately 2.6×10^6 km^3 of ice. Most of the ice sheet receives an excess of snow accumulation over the amount of ice lost to wind, meltwater run-off or other ablative processes. The majority of mass loss occurs at the margin of the ice sheet as either surface melt, which flows into the sea or calving of icebergs from the tongues of outlet glaciers. Many estimates of the magnitude of these processes have been published. The following table (from Ambach, 1980) summarizes an average of five published estimates:

	Accumulation	Surface Melting	Calving	Net
Area (km^2)	1.5×10^6	0.25×10^6		
Average Value (mm/a)	333	-1160		
Mass Flux (km^3/a)	500	-290	-200	10

If these estimates are correct, then this table shows that the Greenland Ice Sheet is in approximate equilibrium and contributes 490 km^3/a of fresh water to the North Atlantic and Arctic Oceans.

How would these contributions change in a different climate? At present, the altitude of the boundary between the accumulation and ablation areas (called the equilibrium line altitude, or ELA) is approximately 1,500 meters above sea level. At Camp Century (76.5°N) the increase in average surface temperature is 1.12°C per 100 meters of elevation (Benson, 1962), while further south along the EGIG line (~70.5°N), this value is 0.6°C/100 meters. Thus, as a rough approximation, one can say that a temperature increase from 3 to 5°C would cause the ELA to increase about 500 meters in elevation. This scenario also roughly corresponds to the predicted atmospheric warming caused by a doubled amount of CO_2 in the atmosphere. In this case, the altered amounts of positive and negative mass flux (assuming no change in the average accumulation and melting rates) are as follows:

	Accumulation	Surface Melting	Calving	Net
Change in Area (km^2)	-0.34×10^6	0.34×10^6	0	0
Change in Mass Flux (km^3/a)	-113	-394	0	-507
New Mass Flux (km^3/a)	+387	-684	-200	-497

Thus, the amount of meltwater discharged into the ocean would increase by 394 km^3/a (or 136%), and the flux of fresh water into the seas would increase to 884 km^3/a. If this temperature change occurred instantaneously, only about half of this increase would reach the oceans; the other half would be refrozen in the near-surface layers of snow in the newly created ablation regions. Within a decade or so, however, this meltwater contribution could be expected to reach the sea.

How would the flow of the ice sheet be affected? This is difficult to say. The mass deficit of 497 km^3/a would cause a thinning rate averaged over the entire surface of 0.29 m/a. Actually, this mass loss would be strongly concentrated at the margins. To return to balance, the ablation area would need to decrease to 0.16×10^6 km^2 (assuming everything else remained constant). The average retreat of the ice edge would be 32 km. These changes in the ice sheet profile would lead to a steeper surface profile, which might imply faster flow and increased calving rates, but the magnitude of these effects are very difficult to assess.

How can remote sensing be used to study the ice sheet? In particular, space-based altimetry, passive microwave measurements and high-resolution imaging hold the greatest utility for application to the ice sheet. Altimetry surveys can provide measurements of elevation change over the entire ice sheet. Repeated every 5 to 10 years, regions of major changes would be readily identified. At the margin, where altimeters have a more difficult time tracking the steeper slopes, imaging systems can provide the best information on the position of the ice edge and its fluctuations. The great difference in microwave emissivity for wet and dry snow provides an obvious opportunity to monitor the extent and duration of the melt region over the entire ice sheet. Some of these studies are presently underway, but more can be done to develop these techniques and emphasis must be placed on the need for future remote sensors to provide repeated measurements.

CLIMATIC IMPACT OF A CHANGE IN NORTH ATLANTIC DEEP WATER FORMATION

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The response of the ocean to climate changes is one of the most uncertain questions regarding the impact of increasing CO_2 on climate and society. North Atlantic deep water formation apparently depends on a complex confluence of different water masses originating in different areas, all of which will presumably be affected by changes in wind, evaporation, etc. as the atmosphere warms. To analyze from first principles what the effect will be on NADW formation is a task which requires an ocean modeling capability not yet available.

As a substitute, one can investigate past climates and see if there is any evidence for alterations in NADW formation. In addition, one can then explore the possible impact of such changes on climate. Such a study allows for an estimate of NADW sensitivity (at least in the past) and of the climatic consequences.

As reported by Ruddiman and Mix, reconstructions of the North Atlantic surface water temperatures indicate a substantial cooling between 11,000 and 10,000 years B.P. Were NADW formation to have ceased, it would have resulted in cooler surface waters; whether the reconstructed temperatures were due to this or some other effect cannot be determined at this time. Nevertheless, it was decided that it would be useful to see what the effect these colder temperatures would have had on the climate.

In a joint study involving Drs. Ruddiman, McIntyre, Broecker and Alan Mix from Lamont and Drs. Rind, Peteet, Russell and Hansen from GISS, a GISS GCM run was made with the North Atlantic surface water temperatures reduced to the Ice Age value from approximately 30-70°N, values consistent with the 11 to 10k B.P. temperature reconstructions. Sea ice was kept at current values, as there is no direct evidence for sea ice changes during this interval. The model was integrated with all other parameters initially identical to those for the current climate, and run for several years. The change in surface air temperature between the second year of the experiment and a five year control run of the current climate is shown in Fig. 1. Strong cooling is indicated downwind from the Atlantic, maximizing over Britian and western Europe, with some effect seen as far inland as central Asia. Little effect is seen upstream, over North

How does this result compare with land evidence of climate change at 11 to 10k B.P.? This time period is known as the Younger Dryas, for the widespread sudden cooling noted in land evidence (ice indications and pollen) over Western Europe. Fig. 2, prepared by Dr. Peteet, shows the palynological evidence for this cooling; it is extremely abundant in Western Europe, and less so as one progresses eastward. Fig. 3 shows that there is little or no evidence for

simultaneous cooling over North America. Thus the temperature change produced in the model by cooler Atlantic Ocean temperatures, as in Fig. 1 is consistent with land evidence for the patterns of actual temperature change. As noted earlier, it is still uncertain whether the surface water cooling was associated with changes in NADW formation.

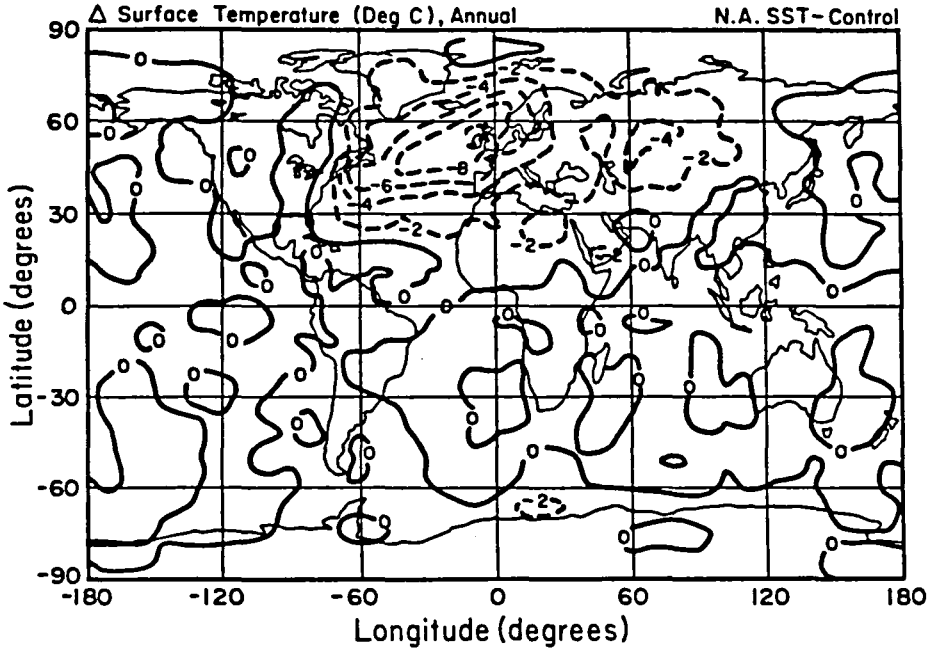

Fig. 1. Change in annual average atmospheric surface air temperature between the experiment with cold North Atlantic surface waters and the control run.

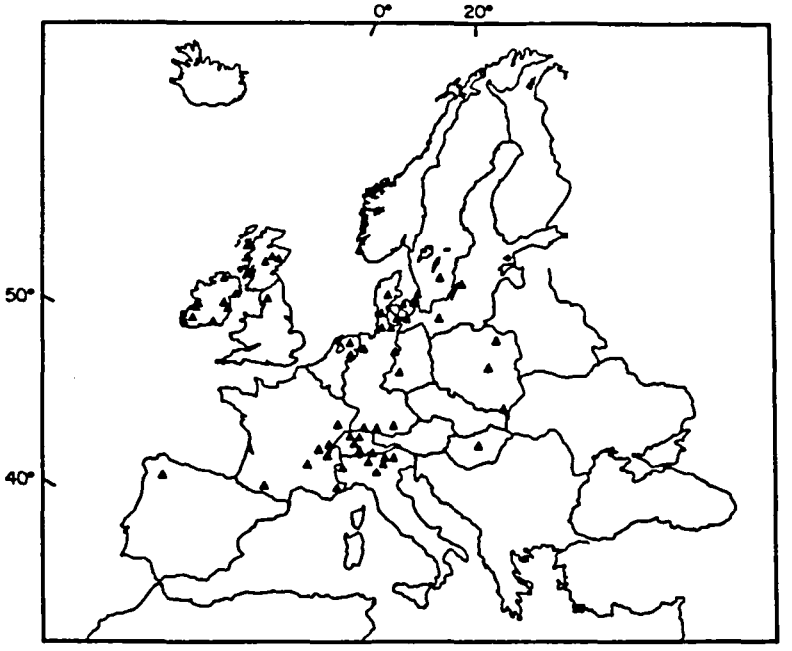

Fig. 2. Triangles indicate locations where palynological evidence for climatic cooling between 11,000 and 10,000 yr B.P., i.e., the Younger Dryas event, has been found.

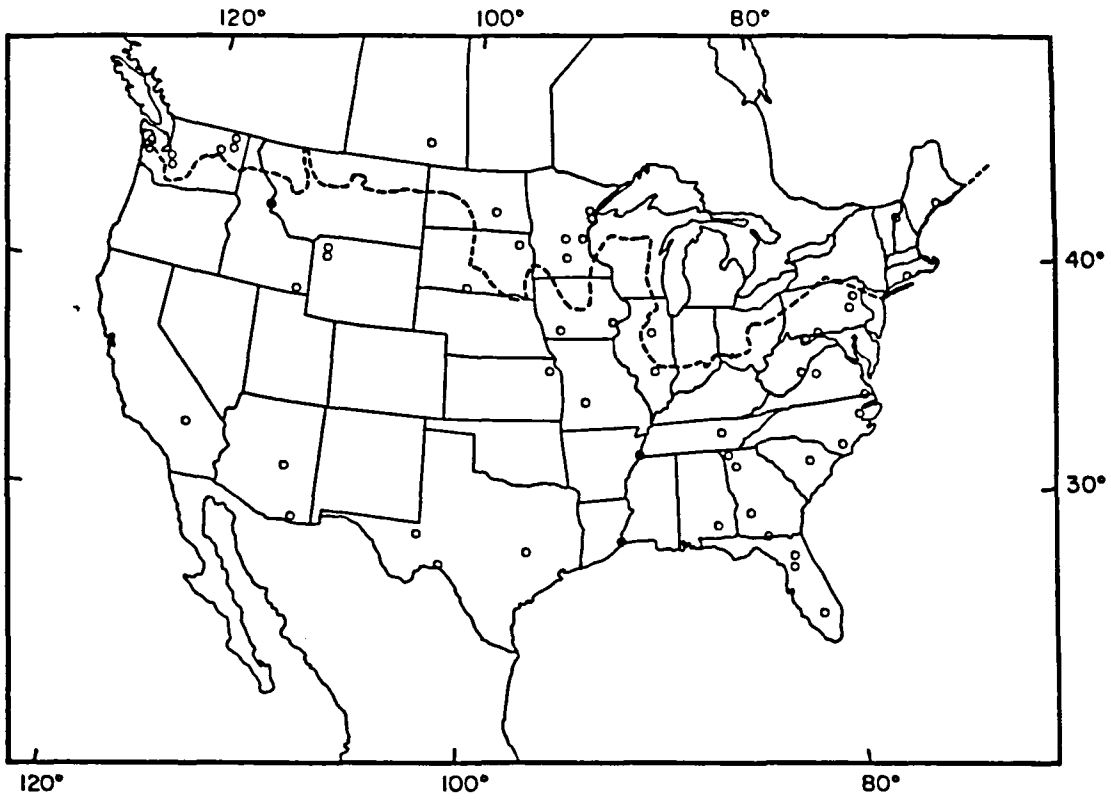

Fig. 3. Circles indicate locations where palynological evidence for climatic cooling between 11,000 and 10,000 yr B.P., i.e., the Younger Dryas event, has not been found.

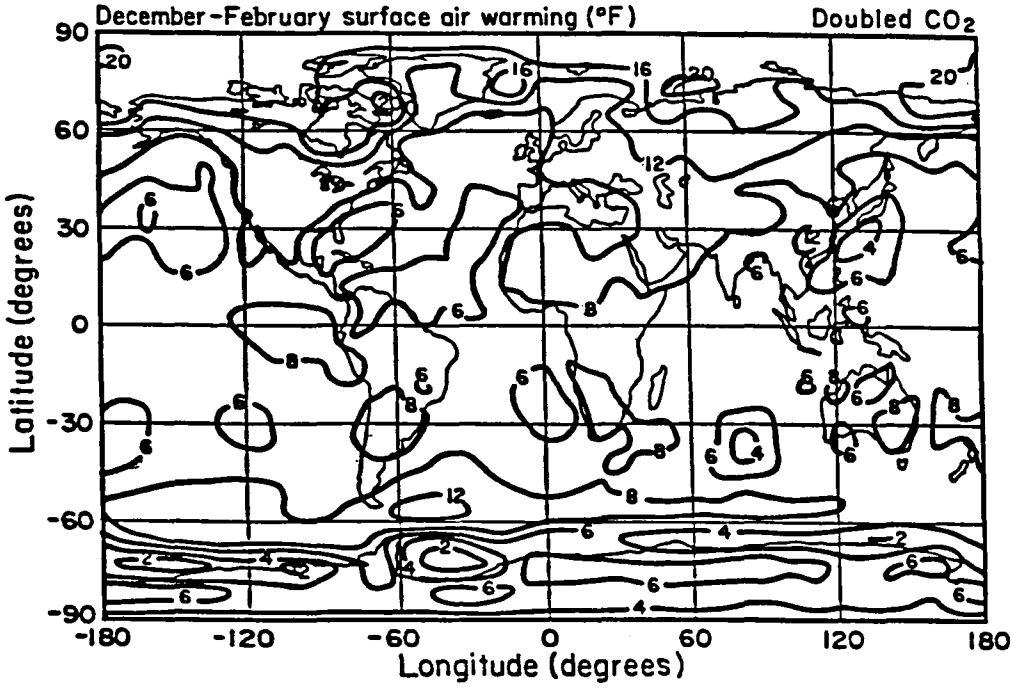

Fig. 4. Change in atmospheric surface air temperature during winter between the experiment with doubled CO_2 and the control run.

What are the implications for the future? Fig. 4 shows the change in surface air temperature produced in the GISS GCM by doubling the atmospheric CO_2, for the months of December through February. Warming of 6-16°F occurs throughout the North Atlantic. Fig. 5 shows the percent change in ocean ice for the same period. Reductions of up to 80% occur in the ice covered areas. The impact of such a large magnitude changes on NADW formation cannot be determined at this stage of our knowledge, but it conceivably could be large. Would this lead to cooler ocean surface temperatures, as a negative feedback, and result in temperature influences such as that shown in Fig. 1? If so, it would obviously modify the regional changes given in Fig. 4, and have further feedbacks on NADW formation. The potential for NADW formation to act as a feedback in a climate change scenario, and thus further influence the climate is a complex problem that needs further investigation.

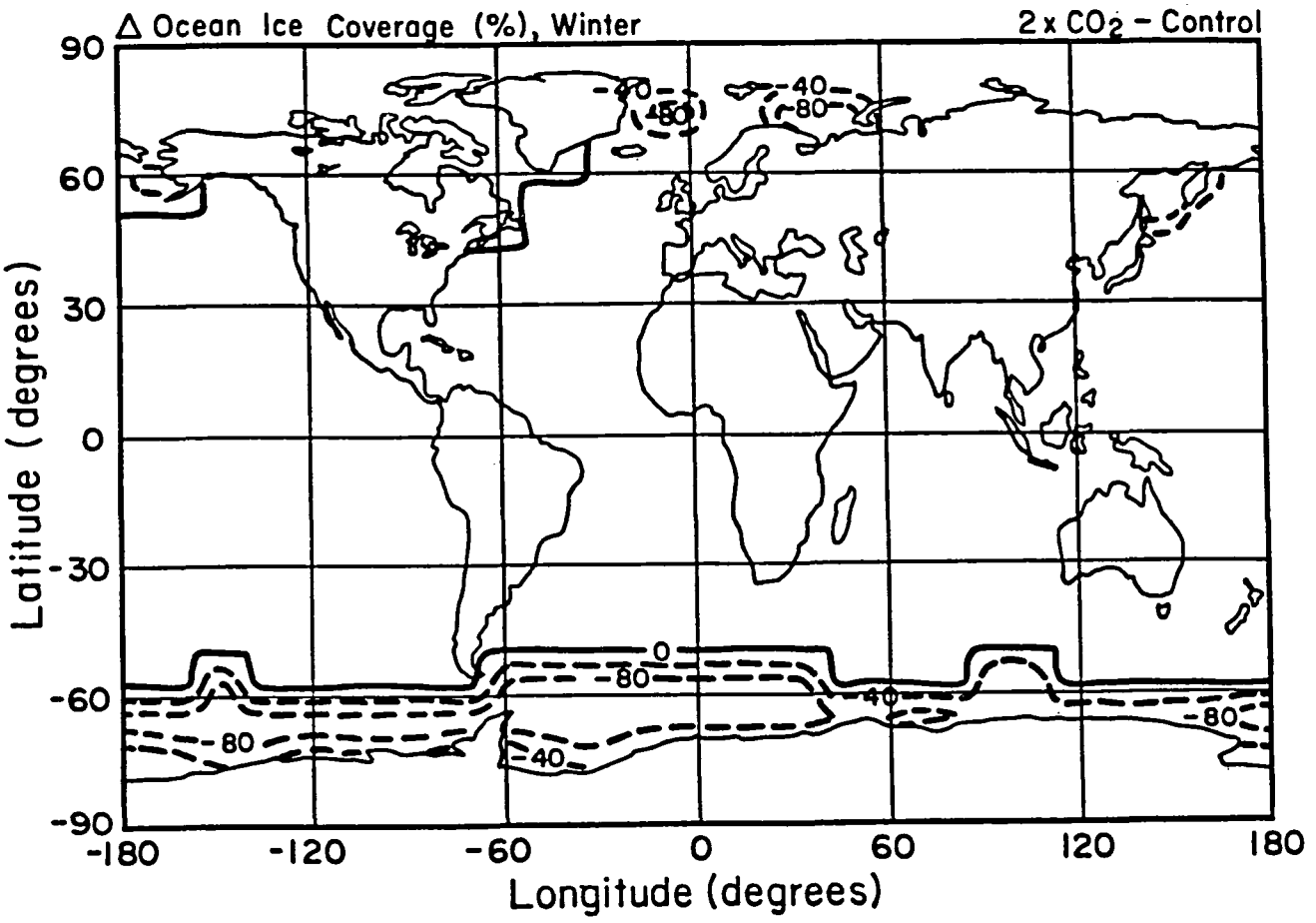

Fig. 5. Percent change in ocean ice during winter in the doubled CO_2 climate.

ORIGIN OF THE GREENLAND ICE SHEET

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A climate model resolving the seasonal cycle and the two horizontal dimensions has been developed at GSFC over the last few years and applied to several problems of current interest. Models of this type are useful when for various reasons a general circulation model experiment is not warranted or not feasible. For example, in cases where the signal to natural variability is small it may be advantageous to first consider such a statistical dynamical model because extremely long runs may be necessary in the application. In this case the simpler statistical dynamical model serves as a pilot study device.

The model developed at GSFC is a thermodynamic model whose solution yields the equilibrium seasonal cycle for the surface temperature field over the globe. The model is essentially a statement of the conservation of heat energy for individual columns of the earth atmosphere system. Various terms such as the infrared radiation flux to space are parameterized with earth radiation budget data from satellites such as Nimbus 6. The primary agent modulating the seasonal cycle amplitude is the heat capacity per unit area which is a strong function of surface type -- ocean surface can store 60 times more heat per unit time than land. By adjusting its few empirical parameters the model can be brought into remarkable agreement with the observed seasonal cycle. The model is described in detail elsewhere (North et al., 1983).

The model is then very useful for looking at the dependence of the seasonal cycle of the temperature on such externally defined variables as the earth's orbital elements (eccentricity, tilt, precession of equinoxes) or the configuration of land-sea geography which can be changed by continental drift. Geological observations imply that the first effect appears to be the cause of the major periodic glaciations which have occurred over the last few million years. Studies with the GSFC energy balance model support this hypothesis further by suggesting the growth of large continental ice sheets especially in North America when orbital elements favor cooler summers which was the case about 125 thousand years ago when the last great glaciation began (North et al., 1983).

Similarly, the model can be used to investigate the seasonal cycle tens of millions of years ago. This study suggests that the configuration of land-sea distribution was such as to not favor ice sheets until about 30 or 40 million years ago, at about which time Greenland and Antarctica became glaciated, although other factors clearly were involved (Crowley et al., 1984). The underlying reason for the transition from an ice-free to an ice-covered Greenland was the warmer summer temperatures caused by the greater continentality in the North Atlantic 50 or 60 million years ago. As the Norwegian Sea

opened and widened just after this time, combined with the separation of Greenland from North America and the simultaneous movement of the North pole towards Greenland, conditions began to favor a continental ice sheet as follows. The presence of more open water in the far North Atlantic moderated the summers enough for snow deposited over the winter to last over the summer months and build up an ice sheet. Fig. 1 (from North and Crowley, 1984) shows the late summer surface temperature distribution computed by the model. Of particular interest is the minimum located in the Norwegian Sea and covering Greenland. Model runs show that this minimum develops as the Norwegian Sea opens. The presence of such a minimum whose minimum value is near freezing can be shown theoretically to induce a transition to a large ice sheet (North, 1984).

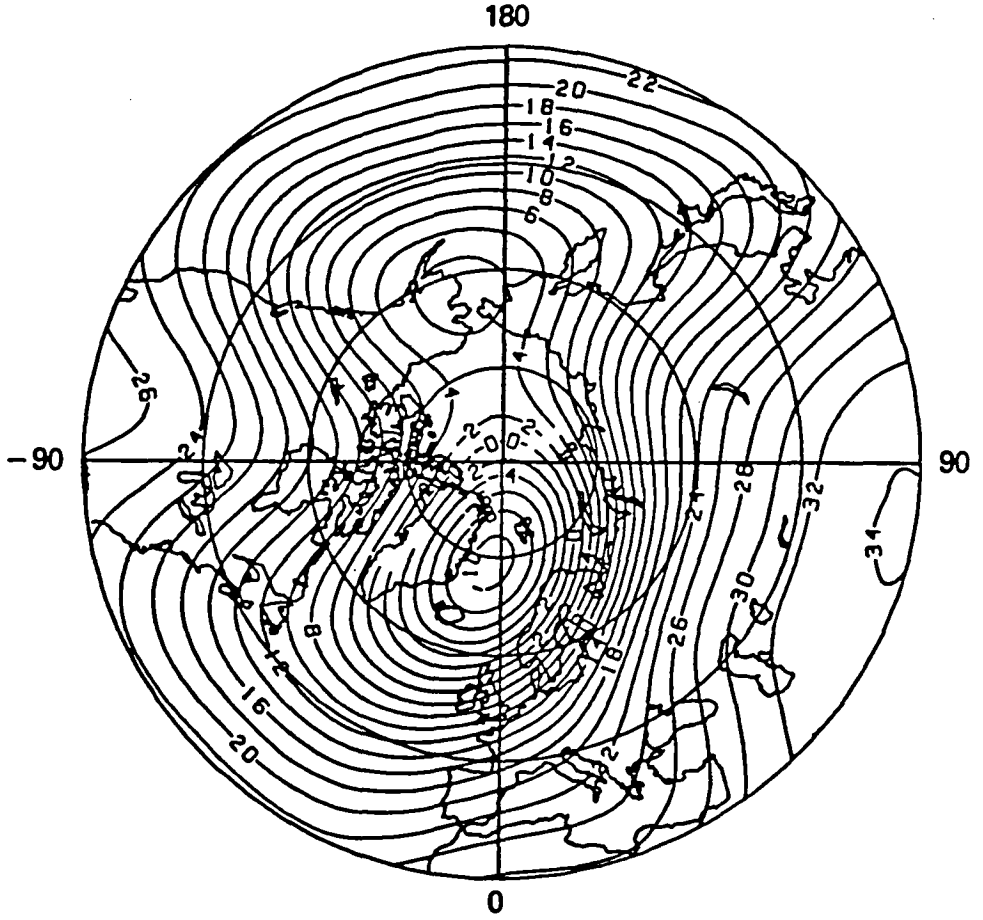

Fig. 1. Polar projection of the July surface temperature distribution as computed by the model, which does not employ an increased albedo over the Greenland ice sheet. In other words the minimum over the Norwegian Sea is induced solely by land-sea configuration.

The model is so crude, not even including a moisture budget, that it would be unwise to consider the theory as conclusive but rather as suggestive of experiments to be done with more comprehensive climate models of the future. This is a formidable task since general circulation models do not yet model the moisture/precipitation cycle with any degree of reliability.

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RECOMMENDATIONS

Oceanographic Measurements

A program to study the formation of North Atlantic Deep Water (NADW) should include an observational strategy which provides a measure of the amount of deep water formed each year and its temperature/salinity (T/S) properties. Also the oceanographic and atmospheric parameters which affect deep water formation processes must be monitored over the same time period. These processes are likely to exhibit considerable interannual variability; therefore, the measurement program should continue for at least 10 years.

Production of NADW is believed to occur in the Norwegian, Greenland and Labrador Seas. However, it does not become NADW until it is able to escape over the 'sill' formed by the Greenland-Iceland-Faeroes-Scotland ridges into the North Atlantic. Our proposed strategy for minimum observations would include annual hydrographic cruises into the region of deep water formation and monitoring of the outflows across the sill. In addition, it is desirable to monitor the several inflows into the region of deep water formation, because each of these sources may have its own variability and sensitivity to environmental factors. An alternative to monitoring these sources may be provided by annual measurements of transient tracers and nutrients, which provide a mechanism to help identify the origin of different water masses.

Hydrographic cruises each April/May into the Labrador Sea, Greenland Sea and Iceland Sea using high accuracy CTD (conductivity/temperature/salinity) systems would provide estimates of the amount and properties of NADW formed each winter. Existing data sets could be analyzed to determine the minimum station spacing required for this purpose.

It is important also to monitor the outflows across the Greenland-Iceland-Faeroes-Scotland ridges. An overflow experiment conducted in 1973 and subsequent programs indicate that the mass flux and the T/S properties can be monitored with existing mooring and current meter technology, although the accuracy, precision and long term stability of moored salinity sensors is still less than desired. While a proper design study would have to be made, it seems likely that such a moored array would need to include of the order of 10-12 moorings with 40-50 current meters equipped with temperature, conductivity and pressure sensors. T/S chains would also be valuable, if these can be developed.

Monitoring of the outflows of deep water to the North Atlantic is not a sufficient measurement strategy by itself, because it does not allow one to distinguish between mechanisms which transform the source water masses and mechanisms which control the blending together of the source waters at the sills. It is therefore necessary to include a survey of the properties of the source waters in the regions of deep water formation each spring, as described above.

In addition, over the same period, it is desirable to monitor the waters entering the formation regions. This would require monitoring the inflows from the Arctic as well as the inflows from the subtropical gyre, and the inflows of intermediate waters from the Labrador, Irminger and Mediterranean Seas. The flows from the Arctic in fact will be monitored, assuming that a proposed multi-year observational program centered on the Fram Strait goes forward as planned. If a similar monitoring of the inflow/outflow across the Greenland-Scotland-Norway ridges were put in place at the same time, this would create an excellent data set from which the processes of water mass transformation in the entire Norwegian/Greenland/Iceland Seas could be modeled and parameterized.

Such a monitoring of inflows from the Arctic is not easy with present technology. Volume transport can probably be measured beneath sea ice using bottom mounted doppler or correlation sonar packages. These instruments are just now becoming available. These instruments might also be further modified to allow the pycnocline depth to be inferred in a manner similar to that of an inverted echo sounder. Monitoring of the T/S of the new surface waters of the East Greenland Current in the presence of ice will be very difficult. T/S chains could be deployed through holes in large ice floes and interrogated and tracked by satellites; however, they would have to be reseeded every few months. Subsurface moored instrumentation would have to be below the keel depths of pressure ridges and perhaps icebergs. The development of a self-contained bottom mounted wired/deck unit/recorder package which would take a profile of the water column at regular intervals using a S/T/P set up to a few meters of the bottom of the ice would be a useful component of this program.

Recovery of moored instrumentation is expensive and difficult in remote and ice covered areas. On the other hand, surface instrumentation moored to the bottom is unlikely to survive a winter season. Consideration should be given to the modification of a package such as used in Rossby's 'RAFOS' floats which would record the data from the instrumentation in some reduced form (perhaps daily averages), return to the surface at some preset time and transmit the entire data via satellite over a period of several days. Such packages could be equipped with sensors so that they would begin transmission only when they were actually on the surface. Packages surfacing under ice could remain silent until they eventually surface in a lead between floes.

In a few areas such as the center of the Greenland Sea and the Labrador and Irminger Seas, satellite drifting buoys with long (<2000m) T or T/S chains might remain quasistationary for periods of months and thus provide oceanographic time series on the evolution of water masses in these basins. Such time series will be a valuable addition to the few oceanographic series still being collected by weatherships.

Studies of the Labrador Sea (Clarke and Gascard, 1983) have shown the importance of episodic events in deep water formation, with the occurrence of deep water formation apparently dependent on the presence of sea ice and strong cooling winds in the appropriate regions. There is a need to develop a better

understanding of such air-sea interaction events, including the role of cyclogenesis processes. This will require data on synoptic scale atmospheric behavior, wind stress and sea ice. Appropriate monitoring appears to be possible with the combination of existing weather stations and planned satellite measurements (see below).

Analysis of convective events and their dependence on large scale atmosphere and ocean conditions would be aided greatly by focused observations of specific events. This could be accomplished by having equipment on the ready and, when available data suggested initiation of an event, sending a plane to release expendable bathythermographs (XBTs) into the indicated region. We understand that a proposal to NSF for such an undertaking is in preparation.

The oceanographic measurement program should be coordinated through the Inter-governmental Council for the Exploration of Sea Ice (ICES) hydrographic committee. Many European nations already have regular oceanographic cruises in these waters for fisheries investigations. Hardware support in the form of high precision CTD systems and technical support for standardization would allow hydrographic data of the required precision to be collected. ICES Service Headquarters may also agree to serve as a data center, and simple mooring packages could be deployed by such fisheries research vessels. The initial need is for the design and package development programs, so that such an oceanographic measurement program can be started in the late 1980s.

In summary, the needed oceanographic studies are:

1) Hydrographic cruises each spring into the Norwegian, Greenland and Labrador Seas to determine the amount and properties of NADW formed each winter for a 10 year period.

2) Monitoring of the meteorological conditions before and during deep water formation to allow investigation of air-sea interaction events, such as the role of cyclogenesis in deep water formation, and monitoring of the sea ice distribution.

3) Monitoring over the 10 year period of the currents and T/S properties of the outflows across the Greenland-Iceland-Faeroes-Scotland ridges into the North Atlantic Ocean.

4) It is also desirable to monitor over this 10 year period currents and T/S properties of the inflows of source waters to the deep water formation regions from the Arctic, the subtropical gyre, and intermediate waters from the Labrador, Irminger and Mediterranean Seas. An alternative is to define a sampling strategy for transient tracers and nutrients with a frequency adequate to define the contributions from different source waters.

5) Development of the required improved capabilities for moored and drifting buoys, which can communicate results via satellite.

6) Release of a network of XBTs into regions of developing deep convection.

Paleoclimatic Research

Climatic records for the North Atlantic and its adjacent lands are to be found in marine sediments, the Greenland ice cap, in marsh and lake sediments and in trees. Each type of record has its peculiar advantages and disadvantages.

The most dramatic changes seen in the ocean, ice and pollen records for this region are pronounced warm events which punctuated glacial time. Oeschger associates these changes with flips between quasi-stable modes of operation of the climatic system. Broecker et al. (1984) extend this idea and associate the rapid warmings with brief reinitiations of the glacially dormant deep water production in the North Atlantic. Much could be done to improve the data base. Regional information exists to date for only the last of these events (i.e., the so called Allerod-Younger Dryas oscillation). It appears in the Greenland ice cap, in high deposition deep sea sediments from the Northern Atlantic, and in marsh and lake sediments from Nova Scotia and western Europe. It is absent in the eastern United States. Similar studies are needed for the events in the 20,000 to 40,000 year time period where several such "Oeschger events" are found in the Greenland ice record.

The Greenland ice cap record shows that these Oeschger events are found not only in the $\delta^{18}O$ (i.e., air temperature record) but also in the dust, ^{10}Be chemistry and CO_2 records. The cause of these changes must be explored. The validity of the ice core CO_2 results as indicators of atmospheric change must be tested through studies of ice cores from Antarctica. While the changes in other properties may prove to be regional, any CO_2 changes must be global.

Trees offer a year by year record of tree growth which extends back several hundred years. To date, few attempts have been made to tie ring thickness fluctuations to historic climate records for Iceland and Scandanavia or to fluctuations in ice accumulation or $\delta^{18}O$ variation in the Greenland and Iceland ice caps. Such studies must be conducted to see if any regional signal can be found which might be related to variations in the intensity of deep water formation.

Unfortunately, our knowledge of historic changes in deep water formation is virtually nonexistent. This makes the job of connecting climate change and deep water production changes very difficult. Nevertheless, a joint study of historic climate, tree ring, and ice cap data should be conducted. In order to do this, the tree ring data especially will have to be augmented.

In summary, the needed paleoclimatic studies are:

1) Exploration for marshes and lakes with pollen records extending back at least 40,000 years, and detailed studies of such records.

2) Retrieval and study of ocean cores from areas of high deposition rate (i.e., >6 cm/10^3 years) in the North Atlantic.

3) Retrieval and study of ice cores from the Crete site in Greenland (site chosen by NSF for the next drilling effort).

4) Extension of the tree ring record to all geographical regions around the North Atlantic basin.

Satellite Data

Satellites measurements have the potential of assuming several key roles in the development of an understanding of deep water formation. Some of these measurements are already planned, but others must still be developed.

In situ oceanographic measurements, of the type discussed above and in proposed 'cage' studies (Dobson et al., 1982), will provide the most precise measurements of ocean heat transport. However, an additional useful estimate of ocean heat transports can be inferred from the combination of radiation budget measurements at the top of the atmosphere and meteorological measurements of atmospheric heat transports. This procedure has been used for estimating the zonal mean heat transport by the ocean. It is practical to obtain very accurate local radiation budget measurements from satellites with a resolution of about 1000 km. If these were combined with meteorological measurements of winds and temperatures at a comparable resolution, it would be possible to infer the divergence of heat in the ocean at 1000 km resolution. The Earth Radiation Budget Experiment should allow a first attempt at this kind of analysis. It may be possible to also extend the results to a much finer horizontal resolution by combining high resolution satellite visible and infrared imaging with the spectrally more complete and accurate radiation budget data.

Fluxes of momentum and moisture at the ocean surface, in addition to heat discussed above, drive the ocean circulation, including deep water formation. Satellites offer the only possibility for measuring these quantities on a global or oceanwide basis. Measurements of surface wind may be possible with a 'scatterometer', an instrument which has been shown to have promise for this purpose. Techniques for obtaining rainfall rates from microwave measurements are being developed, and it has recently been proposed (Liu, 1984) that evaporation could

also be inferred from microwave measurements. A program to fully exploit the satellite capabilities should have high priority. Of course it is important to maintain the weather station network in and around the North Atlantic Ocean, because its data is needed to infer the energy divergence in the ocean and for spot checks of satellite measurements.

Satellite altimetry measurements of the ocean surface provide a potentially powerful technique for investigating ocean dynamics, and such altimetry will be central to the planned TOPEX mission and World Ocean Circulation Experiment (WOCE). Its principal application to deep water formation will probably be to provide information on the large-scale circulation from a combination of altimetry, scatterometry and known and measured hydrography. The objective would be to compute the net flows of surface water in and deeper water out, thus providing estimates of the rates, and changes in the rates, of water mass conversion. Also, it is possible that altimetry could be used to detect and monitor deep water formation events, based on measurement of the surface water depression which should accompany such events. Because of the episodic nature of deep water formation, it is desirable to have the greater spatial coverage obtainable from radar with side-scan capability, for example, synthetic aperture or multiple beam radar. However, it seems appropriate to first look for deep water formation events with existing real aperture nadir pointing radar. If satellite altimetry is carried out at high latitudes it will also allow the possibility of monitoring the Greenland ice sheet mass balance and thus deduction of corresponding freshwater inputs to the ocean.

Satellite microwave monitoring at high latitudes is needed to reveal the coverage and properties of sea ice. This includes accurate determination of the open water fractions within ice covered areas of thin ice which contribute the most to air-sea exchanges. The sea ice monitoring needs to continue on a long-term basis in order to reveal any trends in the nature and coverage of sea ice which presage or accompany changes in deep water formation.

Finally, the capability of satellites to relay data from automatic oceanographic stations needs to be fully exploited. A strong high-technology effort should be made to develop the capability to make a broader range of high-precision measurements from buoys. Such capability is probably the best hope for obtaining adequate oceanographic monitoring on an oceanwide or global basis.

In summary, the principal needs related to satellites are:

1) Monitoring of quantities required to infer air-sea fluxes, specifically radiation budget, sea state (for surface winds) and sea ice.

2) Repeated altimetry of the North Atlantic to help monitor the large scale circulation and to detect and monitor surface water depression associated with deep water formation events.

3) Development and use of greater capabilities for high precision measurements from moored and free-floating buoys with satellite data relay.

Modeling

We believe there is a strong rationale for a focused effort at modeling the North Atlantic ocean/climate system. In order to include the various processes of importance to deep water formation, the ocean, ice and atmosphere must all be modeled at a sufficiently fine resolution. This makes a global model unrealistic at the present time. However, the North Atlantic ocean volume is only about one tenth that of the world ocean, so it should be practical to use high spatial resolution for the North Atlantic with the most powerful existing computers. The North Atlantic is also relatively well observed, as is the climate of neighboring continental regions. Furthermore, the North Atlantic contains the physical processes of interest, including intermediate and deep water formation, sea ice and other high latitude processes, and large fresh water and Mediterranean saltwater inputs. Because of the diverse ocean processes represented in the North Atlantic, it should be possible to generalize successful modeling approaches to the global ocean.

Modeling of the North Atlantic ocean per se can proceed along a number of different approaches, as discussed in the presentation by Semtner. Ideally, successful modeling will include as outputs the rate and locations of deepwater formation and its primary characteristics (such as temperature, salinity, and chemical constituents) plus the modifications of these characteristics as the water descends.

High latitude processes, including sea ice formation, need to be modeled with increasing realism and incorporated into the ocean models. There are indications that sea ice is involved in significant ways in deep water formation, and these processes must be realistically portrayed if we are to be able to correctly project any changes in deepwater formation which may accompany global warming due to carbon dioxide and trace gases.

It is important that modeling be included which realistically portrays air-sea interaction events, such as atmospheric low pressure formation as a result of convective instability, with attendent cyclogenesis in both the atmosphere and ocean. This may require the use of small scale models to explore the detailed characteristics of such events and the relations to deep water formation.

The interaction of atmospheric climate with the ocean (and sea ice) is an ultimate objective of coupled ocean/climate modeling. Although it is essential that the individual model capabilities be improved and tested against today's climate including seasonal and other variations, these models must also be

tested under quite different situations, such as provided by paleoclimate data. In this regard, simulation of the 18K ocean/climate system and rapid transitions such as the Allerod-Younger Dryas may be the most useful of the periods which have the potential for detailed paleoclimate reconstructions.

Finally, we note that the modeling activity for the World Ocean Circulation Experiment (WOCE) includes substantial modeling of the North Atlantic. Our recommendations for modeling are consistent with the WOCE program, but we recommend a strongly focused effort to model the North Atlantic with as much realism as practical for this smaller region. In addition, we recommend that the modeling studies be extended into the Arctic rather than terminating at 65-70°N, to allow adequate study of the deep water formation processes.

In summary, the principal modeling needs are:

1) High resolution modeling of the North Atlantic and Arctic Ocean.

2) Fully coupled ocean/sea ice/atmosphere modeling, which allows analysis of the factors influencing deep water formation.

3) Testing of models against paleoclimate data, as well as against today's climate.

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1. Report No.	2. Government Accession No.	3. Recipient's Catalog No.
NASA CP-2367		

4. Title and Subtitle	5. Report Date
North Atlantic Deep Water Formation	September 1985
	6. Performing Organization Code 640

7. Author(s)	8. Performing Organization Report No.
T. Bennett, W. Broecker, and J. Hansen, Editors	

9. Performing Organization Name and Address	10. Work Unit No.
NASA Goddard Space Flight Center Institute for Space Studies 2880 Broadway New York, NY 10025	11. Contract or Grant No.
	13. Type of Report and Period Covered Conference Publication
12. Sponsoring Agency Name and Address National Aeronautics and Space Administration Washington, DC 20546	14. Sponsoring Agency Code

15. Supplementary Notes

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16. Abstract

This report contains the proceedings and recommendations of a workshop on North Atlantic Deep Water Formation held on June 4-5, 1984 at Lamont-Doherty Geological Observatory, Palisades, New York. The workshop focused on the following questions: What controls the rate of deep water formation in the North Atlantic Ocean and what repercussions would there be from changes in this rate? Strategies for improving our understanding of deep water formation were discussed and recommendations were developed, as summarized in this report.

17. Key Words (Selected by Author(s))	18. Distribution Statement
North Atlantic Deep Water Oceans Climate	Unclassified - Unlimited Subject Category 48

19. Security Classif. (of this report)	20. Security Classif. (of this page)	21. No. of Pages	22. Price
Unclassified	Unclassified	72	A04

For sale by the National Technical Information Service, Springfield, Virginia 22161

NASA-Langley, 1985